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THE UNIVERSITY OF ALBERTA

THE NEAR SURFACE HYDROLOGY OF THE STEVEVILLE BADLANDS, ALBERTA

by

Ⓒ SUSAN E. HOGG

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

OF MASTER OF SCIENCE

DEPARTMENT OF GEOGRAPHY

EDMONTON, ALBERTA

FALL, 1978

Stormy skies over the Steveville Badlands

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled THE NEAR SURFACE HYDROLOGY OF THE STEVEVILLE BADLANDS, ALBERTA

submitted by SUSAN ELSIE HOGG

in partial fulfilment of the requirements for the degree of Master of SCIENCE

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Date July 1. 1978

ABSTRACT

Little attention has been given specifically to the physical hydrology of badlands, particularly their near surface layers, resulting in a lack of understanding of the runoff process in badlands environments. This study examines some of the elements of the near surface hydrology of the Steveville badlands incorporating both classical and more modern hydrologic concepts. Much of the material is not soil, but rather weathered bedrock with high montmorillonite content. Lithologic variations are discussed and their properties relating to hydrologic performance of the lithology such as texture, specific surface, clay mineralogy and shrink-swell phenomena are examined. Precipitation measurements were made in a small basin and soil moisture samples were taken for each rainstorm. Several runoff generation models are proposed for badlands surfaces.

It is concluded that lithology and mineralogy are highly variable which has implications related to the scale of the study. Moisture content of the "soil" is linearly related to rainfall amount below a threshold amount of rain which infiltrates into the "soil." In view of this, the application of partial area concepts to runoff generation in the semiarid environment seems valid. Aspect, the effect of which is dominated by wind, has a significant effect on the moisture content indicating the importance of microclimatic variables in the semiarid environment. No relation could be defined between lithologic variables and moisture content. Structure of the material appears to be more important in runoff generation than does the lithology; surface runoff is more dominant on the sandstone slopes, whereas subsurface runoff is more dominant on the shale slopes. Micro-channel precipitation is an important factor with regard to moisture contents and runoff generation of desiccated materials resulting in a microscale partial area model. Horton's (1945) model does not apply generally in clay badlands; several models apply to the badlands which are represented within one basin. This extends the partial area concept to the microscale in the semiarid badlands environment.

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CHAPTER I

INTRODUCTION

1.1 NATURE OF THE STUDY

Although considerable research has been undertaken in badland environments, (e.g. Schumm, 1956a, 1956b; Parker, 1963; Campbell, 1970, 1974, 1978) little attention has been given specifically to their physical hydrology and in particular the hydrology of the near surface layers. Manifested in this is a lack of understanding of the runoff process in badlands environments.

Classical concepts of surface hydrology are essentially based on the early work of Horton (1945) who outlined the runoff process as being dependent primarily on the rainfall intensity. It was considered that when rainfall intensity exceeded the infiltration rate of a given soil then surface runoff occurred as overland flow. The importance of subsurface flow has been emphasized in more modern studies on runoff generation and the concept of partial area contributions to runoff has been introduced (e.g. Betson, 1964; Kirkby and Chorley, 1967; Dunne and Black, 1970a; Weyman, 1973).

This study examines some elements of near surface hydrology of badlands in the light of both classical and more modern hydrologic concepts. Hydrologic properties pertinent to the runoff process of a range of terrain types present in the Steveville badlands are discussed. Inferences are made about the relative importance of surface and subsurface runoff for each terrain type and possible mechanisms of runoff production are suggested. The value of the partial area concept to runoff, previously regarded as inapplicable in semiarid environments (Kirkby and Chorley, 1967; Freeze, 1972; Garner, 1974; Weyman, 1975), is examined. An attempt is made to outline a physically-based process-response model for runoff for a small basin in the Steveville badlands which should have regional validity and which may be applicable to the more general situation found under badlands conditions wherever they occur. Since much of the material under investigation is not soil, traditional approaches to investigation of hydrologic properties of soils may then prove inappropriate. This study provides information in this regard and aids in the selection of future research methods in badlands environments.

The Steveville badlands provide a striking panorama of deep gullies and steep bluffs with horizontally bedded multi-coloured layers of clay, shale, sandstone, ironstone and coal. This dramatic landscape is characterised by extensive gullying, piping,[1] sharp ridges, residual mounds, hoodoos and smooth micro-pediments typical of badland areas.

As a result of this great diversity in landscape and the different lithologies present, there exist a number of different surface materials which range in type from unvegetated, highly desiccated, disaggregated shale slopes to very smooth, uniform micro-pediments. Lithologies vary considerably but in general they contain large amounts of montmorillonite and consequently much of the material is subject to shrink-swell phenomena. On the areas with a comparatively close cover of vegetation, soils exist. It should be emphasised that no contemporary soil development is present on much of the surficial material in the badlands. This material is essentially weathered bedrock. Soils are restricted to areas where a close cover of vegetation exists. Where soils do occur, profile development is at the juvenile stage. In order to divide surfaces into different categories, a number of terrain types were established. For purposes of this investigation a terrain type is defined as a given area of land as characterised by its surface material present whether it is unvegetated weathered bedrock or soil with a close vegetation cover.

A study of the hydrology of an area usually involves consideration of surface runoff. The interaction of the terrain type with rainfall rather than just rainfall amount is the key factor in the runoff process. Since terrain types vary over an area, particularly with lithology, it follows that various surfaces react hydrologically in different ways to similar rainfall amounts. The Steveville badlands provide a wide variety of terrain types covering a range of lithologies and thus offer an ideal area for the investigation of near surface hydrology in the semiarid environment.

1.2 THE APPROACH

A number of variables are relevant to the hydrological performance of a particular soil. Generally stated these variables include, among others, moisture content or wetness of the soil; the capacity of the soil to serve as a temporary reservoir; vertical and non-vertical movement of water through the unsaturated zone and the entry of water into the soil (Ward, 1975). These variables take on differing degrees of importance in different environments. Those factors affecting these variables, which are most representative of the badlands, will be considered here since these should constitute the major source of explanation for any unique characteristics of the near surface hydrology in the badlands which may become evident.

Factors affecting the moisture content and its variation over the surface include rainfall amount, intensity and duration, terrain type, slope and aspect. Inherent in a description of terrain type is an explanation of the texture of the material, the specific surface and the clay mineralogy present, since these properties govern closely the rate and potential amount of the terrain types moisture holding capacity and hence its efficiency as a runoff generating surface.

In the badlands environment the rate of drying of a soil is an important characteristic affecting the soil as a reservoir. The rate of drying is also related to the rate of desiccation crack formation on the terrain types.

Consideration of the vertical and non-vertical motion of water in the soil involves an examination of the available pathways for water to flow. Desiccation cracking is a striking characteristic of many surfaces in the badlands environments and appears to occur in different degrees on differing terrain types and on some surfaces does not occur at all. An important soil characteristic in this respect is the amount of shrinkage to which a given soil is subject. A large amount of shrinkage implies higher density of relatively wide cracks in the surface than a soil with only a small shrinkage index. Because rainfall is able to penetrate to a greater depth in a cracked soil, the moisture content at runoff threshold level may be higher in a cracked soil than a relatively uncracked one.

Layering of permeable and impermeable layers is also pertinent to the mechanism of runoff production. If water is able to penetrate relatively freely through the soil profile, then the amount of rainfall that is necessary before runoff occurs may be much greater than if a more inhibiting layer exists in the profile. Soil water may build up behind the impermeable layer and a perched water table in the solum or upper weathered material may result thus causing runoff to occur at a much earlier stage of a given storm.

Entry of water into the soil involves consideration of the infiltration process and relative infiltration rates of the various lithologies. Pertinent to this process also, is the amount and degree of desiccation cracking and shrink-swell phenomena of the soil.

The thesis begins with a brief introduction to the problem and the badlands environment. An outline of the approach to the problem and a detailed description of the study area then follows. The context of the arguments presented, which embodies a review of the relevant literature, appears in Chapter III. Field and laboratory methods and methods of analysis are outlined in Chapter IV. This is followed by a discussion of the results and implications thereof leading to the conclusions of the investigation.

Footnote:

[1] Piping is defined by Mears (1963, p. 101) as "... subterranean erosion initiated by percolating waters which remove solid particles from clastic rocks to produce tubular underground conduits."

CHAPTER II

THE STUDY AREA

2.1 LOCATION AND ACCESS

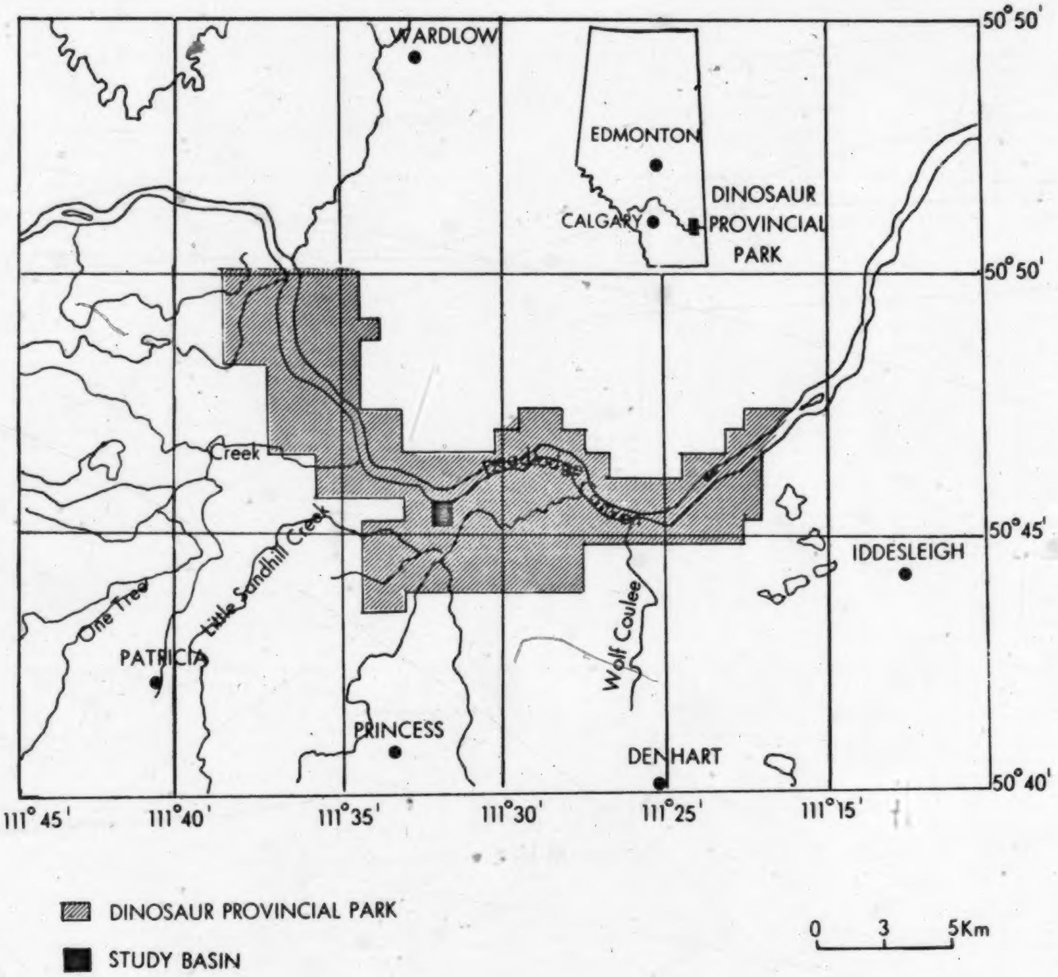
Badlands extend for over 300 km along the Red Deer River between Atlee (near Red Deer) and Nevis in southeastern Alberta (Stelck, 1967). A small drainage basin (0.17 km^2) in Dinosaur Provincial Park, approximately 50 km north of Brooks, was selected for study (Fig. 2.1). The basin is located on the south bank of the Red Deer River near the park entrance with easy access via a service road from the park headquarters.

2.2 GEOLOGY

The badlands are formed in the near horizontally bedded and highly erodible Upper Cretaceous shales and sandstones of the Oldman and Bearpaw formations. The Oldman strata, as defined by Russell and Landes (1940), were deposited in marginal swamps and streams during the withdrawal of the Bearpaw Sea (Byrne and Farvolden, 1959). Dodson (1970) calculated the general dip to be approximately 4.74 metres per kilometre to the northeast. This near horizontal bedding (Fig. 2.2) is a reflection of the deltaic environment in which the sediments were deposited.

The sandstones are frequently characterised by numerous crossbedding features; both channel and overbank deposits are present (Dodson, 1970) indicative of the fluvial nature of their depositional environment. The shales, mostly backwater deposits, "deserve the name of clay rather than shale," (Russell and Landes, 1940, p.62). They are mainly bentonitic, volcanic in origin (Byrne and Farvolden, 1959; Stelck, 1967) and contain

FIG. 2.1 LOCATION MAP


WARDLOW
EDMONTON
CALGARY
DINOSAUR PROVINCIAL PARK
Creek
Little Sandhill Creek
One Tree
PATRICIA
PRINCESS
Wolf Coulee
DENHART
IDDESLEIGH
River
50° 50'
50° 50'
50° 45'
50° 40'
111° 45'
111° 40'
111° 35'
111° 30'
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▨ DINOSAUR PROVINCIAL PARK
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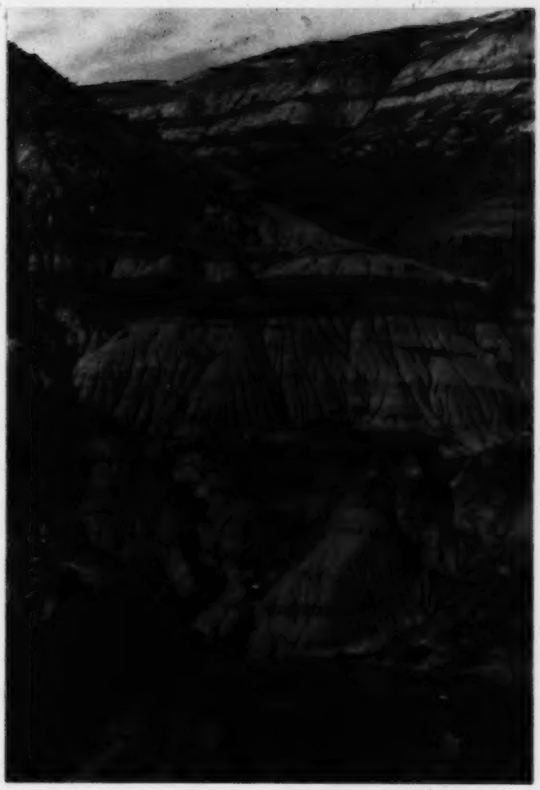


Fig. 2.2 Horizontal bedding of strata of the Oldman formation in the study basin.

considerable amounts of montmorillonite clay minerals (Byrne and Farvolden, 1959). Colour of the bentonite beds ranges from white to grey and yellow-green. The beds are highly fossiliferous containing both fresh-water and marine fossils and the sandstones in particular contain the worlds largest known concentration of dinosaur bones of Upper Cretaceous age, hence giving rise to the name of the park.

At the prairie level and deposited on top of the Cretaceous sediments is a thin layer of Pleistocene till associated with Laurentide glaciation. There appears to be some relationship between thickness of till deposits and the degree of badlands development; the more complex badlands topography being found in areas with only a thin till deposit while little or no badlands topography exists in areas where bedrock is buried by thick deposits of till (Beaty, 1975a).

2.3 TOPOGRAPHY AND LITHOLOGY

The basin is drained by an intermittent stream with three major branches which is incised approximately 100 m below the prairie surface (Fig. 2.3). Badlands dominate the basin topography although their development is subdued in its lower, northern portion. Badlands are defined by Stone (1967, p.215) as:

> "extremely rough topography formed in an advanced stage of gullying in more or less horizontally bedded, poorly consolidated sediments and characterised by sharp-edged ridges separated by narrow and steep gullies."

Slopes are variable in the basin, but topography generally is ultrafine textured (D values average about 250 for the basin).

Two main lithologic units exist in the study area. These are the

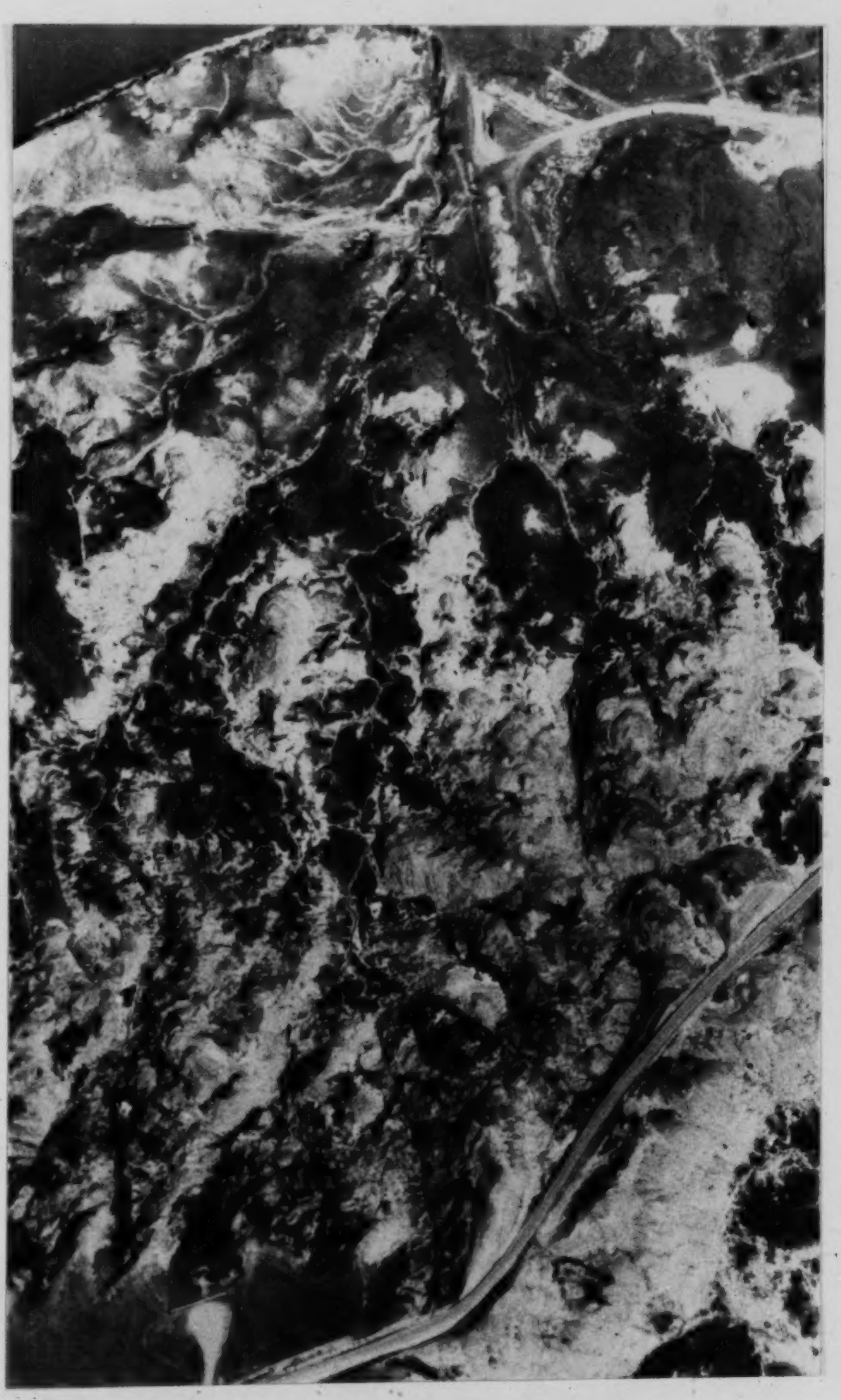


Fig. 2.3 Aerial photograph of basin.

clay-cemented rilled sandstone and the shale. Frequently interrupting the horizontally bedded sandstones and shales are thin deposits of ironstone and well-indurated arkosic sandstone. These sandstones are probably levee deposits (Dodson, 1970; Russell, 1977) in contrast to the channel deposited clay-cemented, rilled sandstone. The layers of ironstone and arkosic sandstone are often weathered and disaggregated, forming armouring debris slopes where they overlie softer shale and sandstone units. Where they occur, these layers exert considerable structural control on both surface and subsurface flow channels. The clay ironstone exhibits classical spheroidal weathering. In localised areas, concentrations of gypsum crystals litter the surface.

Evidence for surface and subsurface erosion is abundant in the form of highly rilled slopes (Fig. 2.4) and considerable pipe development (Fig. 2.5) typical of badlands environments. Rilling is most dominant in the channel sandstone units, being less evident on the shale slopes. This may imply the dominance of surface runoff on sandstone slopes relative to shale slopes. Subsurface pipes in the study area occur in both shale and sandstone units. This raises the question as to the relative significance of subsurface runoff on the two lithologies. Dense networks of rills on the sandstone units are reduced to well spaced more subdued rill networks on the shale units, the transition being abrupt at the sandstone/shale contact. The contact of the two units is also usually associated with a sharp break in slope. Although unproven, a similar, though converse, relationship may exist with respect to pipes, i.e. networks may be more dense in shale but less so

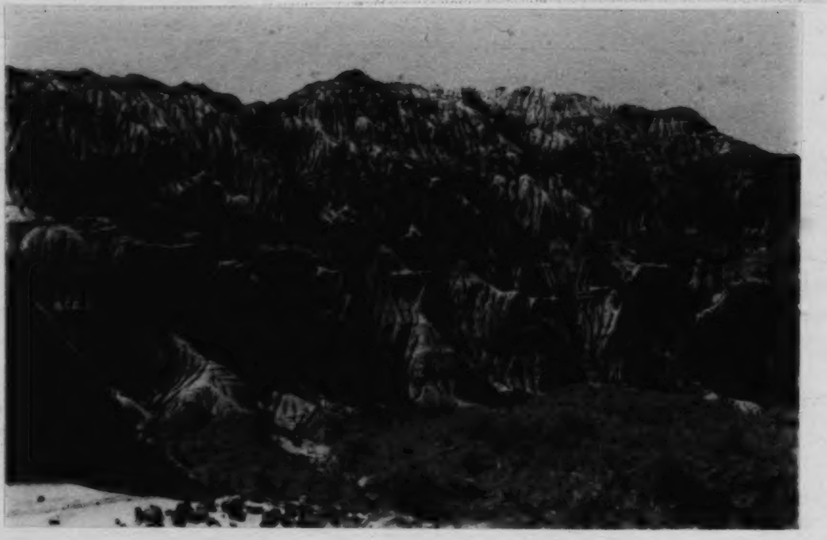


Fig. 2.4 Highly rilled clay cemented sandstone.

Fig. 2.5 Pipe development in the study basin.

in sandstone.

The shale is weathered by desiccation cracking into a loose surface "popcorn" layer which overlies a dense compacted crust of variable thickness above a layer of loose shards (Fig. 2.6). This crust appears to be formed by wetting of the shard layer to such a depth that extensive desiccation cracking does not take place (10-20 cm). The shard layer is an indicator of the volcanic origin of the bentonite beds (Grim, 1968). Individual shards disintegrate immediately on wetting. The crust is impermeable except for occasional structural cracks caused by compaction and is therefore an indication of depth of wetting of the surface material over an extended time period. On interfluves the shale surface is usually convex and unrilled, grading into steep rectilinear slopes on the flanks with small irregular scars (Bryan, et al., 1978). On gentle footslopes, the shale unit is characterised by rill networks although these are not so well developed as on the sandstone unit.

The sandstone has a weathered layer varying in thickness from five to ten millimetres overlying the massive interior in which some non-systematic structural cracks due to consolidation and compaction are present (Fig. 2.7). These cracks provide almost the only pathways for the entry of water into the rock. Desiccation cracks, although scarce, are present on the weathered surfaces and air bubbles, underneath the surface, often produce gaps up to several centimetres wide between the weathered layer and the massive, unaltered rock. The sandstone forms both high and low angle rectilinear slopes (Bryan, et al., 1978) but is more commonly associated with steep concave slopes. The sandstone unit

2.6a. Profile of the "popcorn" terrain. The crust is shown between the two matchsticks. Above the crust is the "popcorn" layer and below is the shard layer.

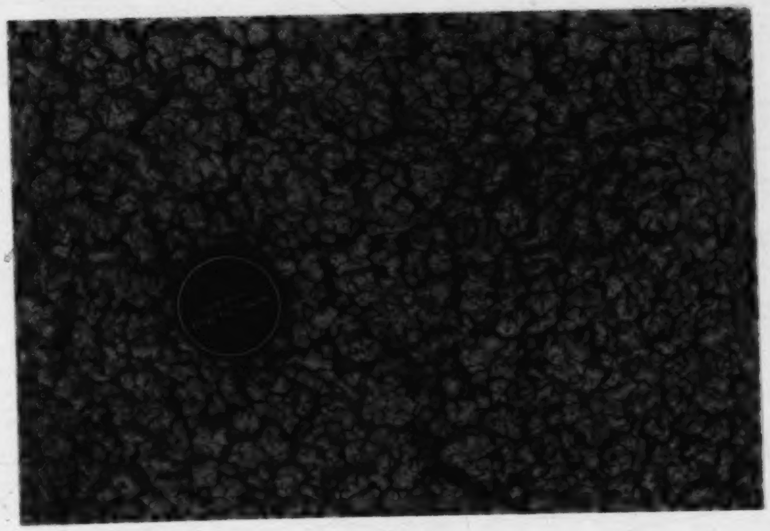


Fig. 2.6b. Surface of the "popcorn" terrain.

Fig. 2.7. Profile of highly rilled clay cemented sandstone. Note the hard impervious rock (dry) beneath the thin (wet) weathered layer.

is characterised by dense rill networks occasionally deeply entrenched (20-50 cm) but more commonly rills are of moderate depth (about 5 cm) and appear uniformly distributed across the slope surface.

Geomorphology of the study basin, as is typical of badlands is complex. Landforms include pipes on small and large scales, remnant bridges resulting from collapsed pipe shafts, debris slopes formed by ironstone and arkosic sandstone particles, residual mounds and hoodoos, and micro-pediments.

Pipes are developed in the study area in bedrock except for one major system developed in alluvial fill at the lower end of the basin. This pipe is similar to those described elsewhere in the literature (e.g. Rubey, 1928; Buckham and Cockfield, 1950; Jones, 1968). Pipes found in the study area vary in size from a few millimetres (Fig. 2.8, microscale pipes) to several metres (Fig. 2.5, macroscale pipes) in diameter. The majority are small and many are collapsed leaving a large number of natural bridges as remnants of their prior existance.

Residual mounds occur throughout the basin where erosion has occurred from all sides leaving an inselberg-like form isolated from other relief. Often weaker materials are protected by more resistant caps producing a mushroom-like landform or hoodoo (Fig. 2.9). Materials forming such caps include rocks, vegetation and fossilised bones.

Micro-pediments are comparable to those described by Smith (1958), Schumm (1964) and Garner (1974) and apparently form at the junction of shale or sandstone and a more resistant layer of ironstone or arkosic sandstone (Fig. 2.10). These gently sloping features were formed by

Fig. 2.8. Microscale pipes with water and sediment flowing out.

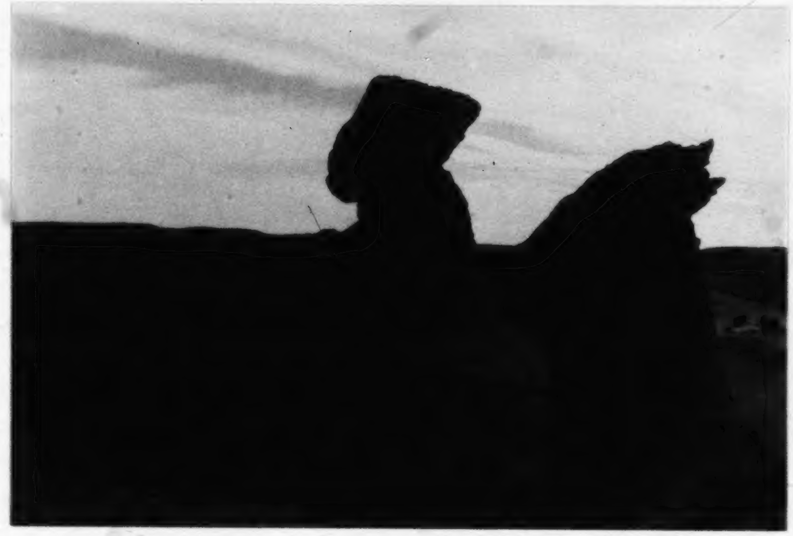


Fig. 2.9. Hoodoos in the study area. The caps are arkosic sandstone.

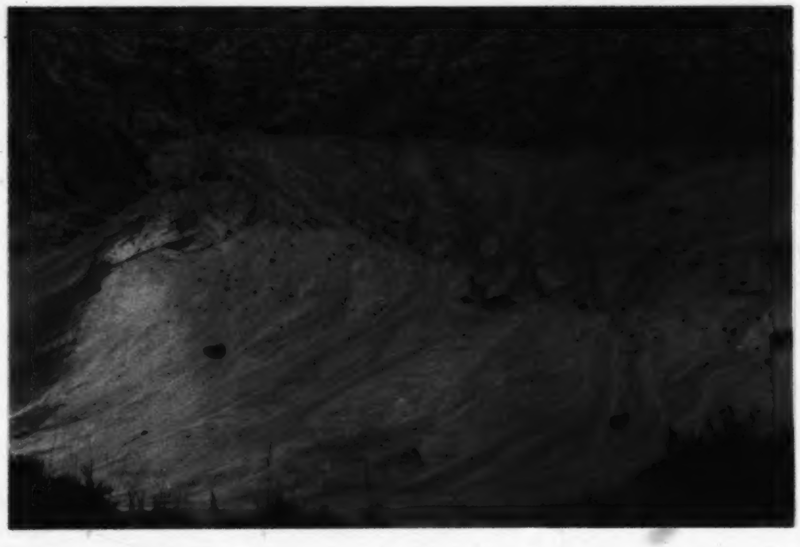


Fig. 2.10. Micro-pediment terrain type. Note the flow lines marking water and sediment issuing from micro-pipes at the base of the shale slope (plot 6). Sheetflow is occurring on the right of the photograph.

repeated alluvial deposition, each layer one millimetre or less thick and frequently containing voids and well-developed vesicles (Bryan, et. al, 1978). The fragile laminae may be easily separated and the surface is very smooth thus allowing surface runoff to occur as a continuous sheet across the surface. These landforms are now essentially surfaces of transport for debris from the slopes they surround (Bryan, et. al., 1978).

2.4 VEGETATION AND SOILS

The study basin is sparsely vegetated in the southern portion where badlands development is most prominent. In the northern section, however, a close cover of xerophytic and mesophytic mosses, grasses and forbs is present, along with cactus species that include prickly pear (*Opuntia spp.*) and barrel cactus (*Cactus mammilaria*) (Fig. 2.11). Sagebrush (*Artemesia spp.*) and greasewood (*Sarcobatus vermiculatus*) grow in areas of moisture concentration such as at pipe inlets and outlets in both the northern and southern areas of the basin. Lichens grow commonly on bare rock and, particularly on the micro-pediments, serve as caps in the formation of miniature hoodoos. Sagebrush and greasewood stems and roots serve a similar function on the almost bare shale and sandstone slopes.

Soil is best developed in the northern part of the basin where the grass and moss cover is most dense. The study area is in the brown soils zone of Alberta (Atlas of Alberta, 1969) but development is very poor. The soils are classified as orthic brown chernozems according to the Canadian System of Soils Classification (1976) and show some evi-

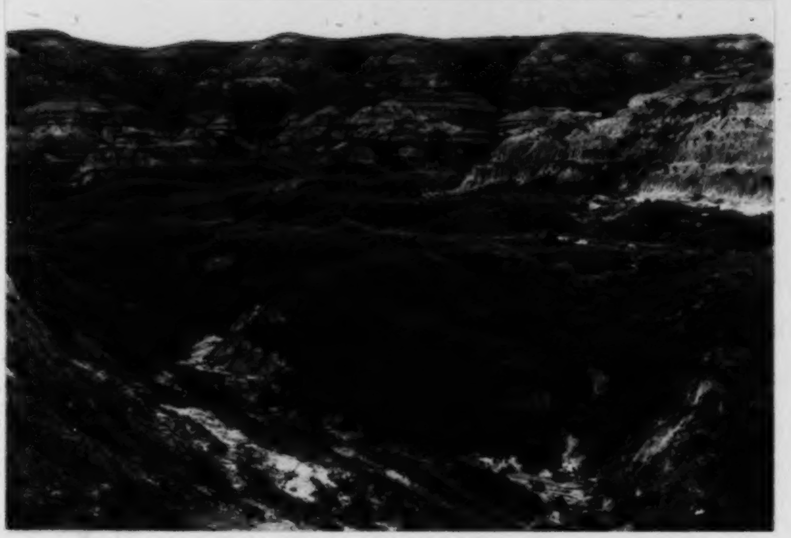


Fig. 2.11. Vegetated area (foreground) of the lower part of the basin. View is looking south towards the unvegetated badland topography of the basin (background).

dence of oxidation as do many soils of arid areas. The "soil" of the remaining sparsely vegetated area in the basin is essentially weathered bedrock and shows no real signs of soil development at all.

2.5 CLIMATE

The study area is in the Prairie Provinces climatic region of Canada and is generally a semiarid area with hot summers (Climate of Canada, 1960). Under the Köppen climatic classification the area has a BSk climate and according to the Meigs (1953) system devised for the UNESCO Arid Lands Research Series, the area is classified as Sb02 (Campbell, 1974). Brooks meteorological station (50 km south of the basin) is representative of the area since over much of southern Alberta the climate is uniform (Longley, 1972), Table 2.1.

Mean annual temperature for Brooks is 3.9°C while the extreme annual minimum is -46.7°C and the extreme annual maximum is 40°C. The mean temperature is less than 0°C for the months November to March (Climate of Canada, 1960), and the average number of days with frost is 199 days.

Mean annual precipitation is 300-400 mm with mean annual rainfall being 240 mm and mean annual snowfall 102 cm. June and July are the high rainfall months with 70 percent of the mean annual precipitation falling in April to September (Campbell, 1970). Precipitation shows extreme variability from year to year with differences between the extreme annual amounts exceeding the mean annual total (Climate of Canada, 1960). The greatest total precipitation received in 24 hours has an annual average of 90 mm (Canadian Normals). The greatest 24 hour

Table 2.1 Climatic data for Brooks, Alberta

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Mean daily temperature (°C)	-13.1	-10.4	- 4.7	5.2	11.6	15.3	19.4	17.5	11.9	6.0	- 3.1	- 8.8	3.9
Mean rainfall (mm)	0.2	0.5	1.3	10.6	38.8	37.4	37.8	52.1	31.0	8.6	1.5	0.7	240.1
Mean snowfall (mm)	167.0	167.0	195.5	100.9	20.3	-	-	-	17.7	91.4	121.9	142.0	1023.7
Mean precipitation (mm)	17.0	17.3	20.8	21.6	40.8	37.4	37.8	52.1	32.7	17.8	13.7	15.5	344.0

total received was 90 mm in June, 1934, while the same statistic for snowfall was 305 mm received in February, 1937 (Canadian Normals). Rainfall intensities on the Prairies, however, are generally much higher than those measured by the sparse Atmospheric Environment Service gauge network (Toogood, 1977) suggesting that these figures are probably underestimating the true value of rainfall intensities. Campbell (1970) estimates that on the average one-tenth of the annual precipitation falls in one day.

Drying winds reduce the effectiveness of precipitation. The prevailing wind direction is southwest with winds from this direction also showing the greatest wind speed (15.1 km/hr). Winds from the northeast and southeast occur only seven and eight percent of the time respectively at speeds of 15.6 km/hr and 14.3 km/hr. Winds from other directions occur 11-16 percent of the time at speeds of 15.3 to 21.2 km/hr (Longley, 1972).

Rainfall usually occurs as fairly intense convective storms which are highly localised within the area to the extent that microclimatic variables play an important role in the rainfall patterns. Aspect is of importance both in terms of the receipt of rainfall from specific directions and also in terms of the rate of desiccation of the slopes. Temperatures in localised basins may be extreme, particularly in areas of light coloured shale slopes and the almost white micro-pediments where the degree of reflectivity between slopes is high. Frost heave may be important in localised areas where moisture is available, such as from a local snow patch. Due to the frequent strong winds and the intermittent nature of the storms, however, desiccation of the slopes is intense and on the whole subsurface moisture is non-existent.

CHAPTER III

CONTEXT AND LITERATURE REVIEW

3.1 INTRODUCTION

In any discussion of runoff processes, two basic concepts are involved: the cyclic movement of water through an open system and the relationship between inputs and outputs of water for any part of whole of the system (Weyman, 1975). This chapter is a discussion of literature relevant to the present study within the context of these two concepts. First, the infiltration process and the runoff process are discussed in classical terms. Since the study area involves swelling and therefore "non-classical" soils, a brief outline of the hydrology of swelling soils follows and shrink-swell phenomena related to soil water storage and movement are discussed. This is followed by a more detailed review of relevant concepts from research in badland areas.

3.2 INFILTRATION

Since the soil surface is the link between the atmosphere and the subsurface parts of the hydrologic cycle (Dooge, 1967), the entry of water into the soil has been a much studied process in hydrology. Although infiltration is strictly a surface phenomenon (Satterlund, 1972) the process can only be properly understood as a function of moisture storage and movement (Ward, 1975). The term infiltration has been defined as the "process of water-entry into the soil, generally (but not necessarily) through the soil surface and vertically downward" (Hillel, 1971, p.131). The infiltration rate is the maximum rate at which a soil in a given condition, at a given time, can absorb rain (Richards and Gardner, 1952). Soil infiltrability is the water flux which the soil profile can absorb through its surface when it is maintained in contact with water at atmospheric pressure (Hillel, 1971). The rate of infiltration may determine the amount of runoff for a given storm. Cumulative infiltration has a curvilinear time dependence with a gradually decreasing slope. Generally the soil infiltrability of a dry soil is high at the beginning of a storm and tends to decrease monotonically and eventually to approach asymptotically a constant rate or steady state infiltrability (Hillel, 1971).

Various infiltration "laws" have been developed to express the rate of infiltration as a function of time elapsed since the inception of surface flooding and, in particular, to account for the rapid decrease from initially very high values and, for uniform soils, the approach to an ultimate constant value (Childs, 1969). Models or laws developed include those of Green and Ampt (1911), Horton (1930, 1933, 1945), Kostiakov (1932), Gardner and Widstoe (cited in Gardner, 1967), and Philip (in a series of papers 1957-1958). Philip's thorough analysis of the problem based on the flow equation showed that the equation of Green and Ampt (1911) was a solution for the flow equation in a special case. Considerable research has been focussed on testing these theories in the field (e.g. Childs, et al., 1957; Rubin, 1966; Gardner, 1967; Smith and Woolhiser, 1971).

Basically two groups of factors control the infiltration rate (Satterlund, 1972). First, factors which determine how rapidly the surface can absorb water and, second, the rate at which the water is applied to the surface. These have been summarised by Musgrave (1958), Parr and Bertrand (1960) and Satterlund (1972):

1. Surface conditions and amount of protection from the rain.
2. Internal character of the soil mass including pore size, depth or thickness of the permeable portion, degree of swelling, content of organic matter and the degree of aggregation.
3. Soil moisture content and degree of saturation.
4. Duration of rainfall.
5. Season of the year and temperature of the soil and water.
6. Effects of micro-organisms.
7. Frost action.

Much of the material in the badlands has an indurated crust 10-20 cm below the surface and it is commonly realised that soils with crusts on their surface, or with layers of permeable and impermeable horizons adjacent to each other do not conform to traditional infiltration theory. Several approaches have been made towards a solution of the flow equation for crust-topped and layered soils (e.g. Hanks and Bowers, 1962; Hillel, 1964). Both the crust and the underlying soil apparently affect infiltrability (Hillel, 1971). This contradicts the generally supposed concept that it is the least permeable horizon alone which controls flow rate in a layered soil. The suction which forms at the interface of the layers is such as to create a gradient through the crust and a conductivity in the transmission zone which results in an equal flux through both layers (Hillel, 1964). Hillel (1964) showed how the presence of a surface crust can shorten the initial (falling rate) period during infiltration, decrease infiltration capacity and affect the moisture content profile of the soil.

A theory has been presented (Hillel and Gardner, 1970) which describes transient infiltration into both uniform and crust-topped profiles of initially dry soil. Three stages of infiltration into crusted profiles were recognised:

1. An initial stage in which the rate is finite and dependent on crust resistance and on effective subsoil suction.
2. An intermediate stage in which cumulative infiltration varies approximately as the square root of time.
3. A later stage in which cumulative infiltration may be expressed as the sum of a steady and a transient term, the latter being negligible for long time periods.

Cumulative infiltration was shown to decrease with increasing crust resistance particularly in coarse textured and coarse structured soils. Cumulative infiltration of crusted profiles is a function of their transmission diffusivities and it follows that infiltration into a crusted profile may be described by the approximation that water enters into the sub crust soil at a nearly constant suction, the magnitude of which is determined by the crust resistance and the hydraulic characteristics of the soil (Hillel, 1971).

3.3 THE RUNOFF PROCESS

> "In 1684 Edmé Mariotte carried out measurements in the Seine river basin that provided the first experimental proof that rainfall is sufficient to sustain river flow. Almost 300

> years later we can argue that very little more about the rainfall-runoff process is known."
>
> (Freeze, 1972, p.1272)

It is now recognised that runoff arises in response to a complex interaction between surface flow and saturated and unsaturated flow areas (Freeze, 1972). Traditionally water seen flooding over the surface or standing in depressions has been regarded as the result of rainfall intensities in excess of infiltration rates (Horton, 1945). This form of runoff is generally known as infiltration-excess overland flow (Calver, et al., 1972). As the infiltration capacity declines during a storm, theoretically the rate of surface water accumulation will increase. Because of the measurement of very high infiltration rates (70-200 mm/hr), and the restriction of surface water to specific areas of slope, evidence suggests that Horton's (1945) concept is incorrect for humid areas (Weyman, 1975).

Much discussion on the significance of subsurface flow has arisen from the early work of Hursch (1944), Berry and Ruxton (1957), Ruxton (1958) and Bunting (1961, 1964). Whipkey (1965) investigated subsurface flow from a piping system under a forested area which has also been discussed by Jones (1971) and the Institute of Hydrology (1971). Hewlett and Hibbert (1965) discussed translatory flow within a soil profile - movement is apparently due to thickening of the water films surrounding solid particles resulting in a water flux pulse as the saturated zone is approached similar to a wetting front advance.

Kirkby and Chorley (1967) show that throughflow is capable of producing storm throughflow hydrographs and they demonstrated that the

Horton (1945) infiltration model of surface runoff and erosion was more limited in geomorphic application than had been previously recognised. Horton's (1945) model was considered to be most applicable to clay badlands which were considered to have low infiltration capacities and little vegetative cover; as such, Horton's (1945) model was viewed as one end member of a wide spectrum of runoff/erosion models. Kirkby and Chorley (1967), using data derived from Whipkey (1965) and Betson (1964), developed the other end member which applies to slopes with high infiltration capacities and relatively thick soil covers where throughflow dominates and overland flow, with its attendant channel initiation only occurs in a few restricted areas. Experimental results of Kirkby and Chorley (1967) were consistent with the infiltration theory of Philip (1957) indicating different response of throughflow to rainfall in the unsaturated upslope area and in the saturated zone at the base of a slope near the channel.

Amerman (1970) found that runoff producing areas were located in seemingly random fashion on ridge tops, valley slopes, and valley bottoms. The runoff production areas were not necessarily connected to the perennial stream by continuous surface flow. Surface runoff was often absorbed before reaching the channel. Hewlett and Hibbert (1965) proposed a conceptual model whose major feature was the contribution to storm flow by development and expansion of saturated zones along valley floors and the lower portions of hillslopes. Ragan (1967) found only a small but variable proportion of the watershed ever contributed to the storm hydrograph.

Working in Sleepers River Catchment, Vermont, Dunne and Black (1970) found infiltration capacities of the soils exceeded rainfall intensities in the vast majority of the storm events. Overland flow as postulated by Horton (1945) was not observed. Runoff hydrographs showed the same features as those in which overland flow was observed. They concluded that under antecedant conditions which prevail throughout most of the year, the magnitude and general shapes of hydrographs from first order drainage areas were controlled by precipitation that fell onto the stream and onto wet areas along and at the head of the stream. As these areas allow virtually no infiltration and little storage, the response of runoff from them has two characteristics: the ratio of runoff to rainfall is close to unity and runoff from these small areas is extremely sensitive to fluctuations of rainfall intensity. Subsurface flow proved to be too small, too late and too insensitive to fluctuation of rainfall intensity to account for the rapid rise and fall of discharge. Dunne and Black (1970, p.1308) suggested that

> "as a description of the major source of storm flow in humid regions, and as an important alternative to the Horton model in such regions, the subsurface stormflow model seems to be less useful than has been claimed."

It was concluded that the dynamic nature of these partial areas may explain the value of an antecedant precipitation index in predicting storm runoff from humid areas where the more familiar relationship between antecedant moisture and infiltration appears irrelevant.

From a theoretical approach Freeze (1972) supported the evidence of Dunne and Black (1970) suggesting stringent limitations on the occurrence of subsurface stormflow as a quantitatively significant runoff component. According to Freeze (1972) subsurface stormflow is a feasible mechanism only on convex hillslopes that feed deeply incised channels and then only when saturated soil hydraulic conductivities are large. On all concave slopes and on convex slopes with low permeabilities, hydrographs are dominated by direct runoff through very short overland flow paths from precipitation on transient near-channel wetlands. On these wetlands surface saturation occurs from below because of rising water tables that are fed by infiltration rather than by lateral subsurface flow. Freeze (1972) concluded that the saturated hydraulic conductivity of the soil exerts far greater influence on the runoff generating system than the properties of the rainfall event or the soil slope configuration. Freeze (1972) maintained that the investigation of surface runoff can be reduced to an examination of the mechanisms that cause lateral flows to first order channels. Apparently a fundamental feature of all but the Horton hydrograph is that storm runoff is only a small percent of the total basin precipitation.

Investigating interflow in mountain forest soils in British Columbia, Chamberlin (1972) found that interflow was dominantly in the form of saturated flow over shallow bedrock, aided by rapid transmission through intervening soil. Unsaturated flow above the phreatic surface was dominantly vertical. Chamberlin (1972) distinguished between a "closed soil" (a soil in which water passes to a water table or stream by means of continuous unsaturated flow) and an "open soil" (a soil in which a large quantity of water passes to the water table or stream channel through pathways which circumvent most of the soil mass).

Accordingly, much of the surficial material of the badlands could be defined as an "open soil." As Chamberlin (1972) pointed out, open and closed soils are identical since physical obstructions or voids may be viewed as elements of the soil matrix whose hydraulic conductivities are either very small or very large. He concluded that open and closed soils may serve as two types of source areas which determine the hydrologic response of a given watershed.

In summary then, alternative mechanisms to Horton's (1945) model have been suggested. These may be viewed as a continuum with infiltration-excess overland flow at one extreme and Kirkby and Chorley's (1967) throughflow model at the other. In between the extremes is a combination overland flow - throughflow model (Betson and Marius, 1969; Dunne and Black, 1970a) where precipitation saturates the soil before runoff occurs. Overland flow resulting from this mechanism is termed saturation - overland flow and is controlled by the soil moisture pattern existing at the beginning of the storm (Weyman, 1975).

Depression storage, which is water held in all the minor surface irregularities, must be satisfied, however, before either mechanism of overland flow can operate. Within a basin the occurrence of overland flow will depend upon the distribution of soil moisture (Weyman, 1975). Generally the importance of overland flow increases as the soil and bedrock become less permeable (Weyman, 1975). Emmett (1970) developed a theoretical model for overland flow on hillslopes and showed flow occurs as laminar near the divide, grading to fully turbulent some distance downslope. Emmett (1970) concluded that slope steepness and the length of each slope facet are controlled by the runoff rate and the initial gradient at the top of the slope. This relates the shape of each profile to its climatic and geologic environment. Hills (1971) found, however, that less than 30 percent of rain events could be expected to produce overland flow and even then only on restricted areas.

The importance of direct channel precipitation has only recently been emphasised. In three small basins with a range of parent materials it was found that a substantial portion of the storm hydrograph generated within a section of the basin was formed by direct channel precipitation (Ragan, 1968; Dunne and Black, 1970b; Weyman, 1973). On a microscale, desiccation cracks in a heavily cracked soil, such as occur in the badlands, may be regarded as micro-channels subject to such direct "channel" precipitation and may form a large proportion of the total surface area.

Calver, _et_ _al_., (1972) described non-linearities in the rainfall-to-runoff relationship and pointed out, as would be expected, that the initiation of overland flow will lead to greatly increased runoff over that in which no overland flow occurs. Multi-layered soils, in which hydraulic conductivity decreases with depth, may have saturated conditions building up at several different horizons under suitable antecedent moisture and rainfall conditions. Also, there are delays between rainfall and effective lateral flow which cause non-linearities in the rainfall-runoff relationship. Again, antecedent moisture conditions help to control this delay. The spatial variability of runoff also contributes to non-linearity. Specific areas of overland flow have been identified and include portions of thin soil A horizons (Betson and

Marius, 1969), downslope parts of long slopes (Jamieson and Peters, 1967) and topographic hollows (Dunne and Black, 1970a). These three sources of non-linearity in the rainfall-runoff relationship result in hillslope runoff which varies areally and over time as a response to input and physiographic controls. "Process-oriented flow models must take all these non-linearities into account" (Calver, et al., 1972, p.199). It is suggested that concave profile and contour areas within a catchment are likely to dominate runoff. This is also implied by Huggett (1975).

In conclusion, Horton's (1945) infiltration-excess overland flow model is generally considered applicable for zones of heavy clay soils and for semiarid areas (e.g. Kirkby and Chorley, 1967; Freeze, 1972; Weyman, 1975). This notion is questionable when the effects of piping desiccation cracking and layering of soils are considered for the semi-arid badlands environment.

3.4 HYDROLOGY OF SWELLING SOILS

Since much of the material of the badlands is subject to shrink-swell phenomena it is necessary to briefly discuss the hydrology of swelling soils. Philip (1969) was one of the first to recognise the major differences between the hydrology of swelling soils and that of nonswelling soils. A review of the hydrology of soils subject to volume change was presented by Philip (1971).

The distribution of hydraulic conductivity is different for a swelling soil than for a rigid soil. Since variations in surface topography cause variations in overburden potential for a swelling soil then surfaces of constant moisture ratio are also affected by topography unlike a nonswelling soil (Philip, 1971). In a nonswelling soil a state of disequilibrium exists when the moisture ratio decreases rapidly away from a water body. Such moisture differentials will exist in a swelling soil even when a true disequilibrium exists (Philip, 1971). In this case the differential persists not because of a lack of hydraulic conductivity but because of a lack of difference in total potential. It has been discovered that steady upward flows may occur against a moisture gradient (Philip, 1971; Sposito, 1975) in swelling soils. This is explained by the overburden pressure which increases with depth and consequently the rate of evaporation can be expected to be increased by the existence of a negative pressure gradient brought on by the overburden (Sposito, 1975).

It has been found experimentally, as Philip (1971) predicted, that infiltration in swelling materials has more in common with capillary rise than with infiltration in a rigid soil (Smiles, 1974). This is because in a saturated swelling soil infiltration is accompanied by a net increase in the gravitational potential energy of the system. For initially saturated clay, it was shown that there is a straight line relation of the height of a sample of clay changing as a function of the square root of time (Smiles, 1974). The change in height of the clay sample was interpreted as being equal to the cumulative volume of infiltrated water. For saturated clays Smiles (1974) demonstrated the effect of gravity is so reduced that for extended periods of time, infiltration may be regarded essentially as a sorption phenomenon. It was concluded that the infiltration process in a swelling soil is an interchange of water downwards with solid upwards. It was also suggested that hysteresis in the diffusivity-moisture ratio relationship may not be as significant in swelling clays as in rigid soils.

Infiltration experiments conducted by R.H. Read (cited in Collis-George, 1977) on open-cracked clays show that there is a very short time effect which corresponds to part of the cracks filling with water. As a consequence of swelling the cracks then close and the mechanisms of sorption and/or transmission become dominant in the infiltration process. Thus by including the integral infiltration, i_I, into the surface cracks ("micro-channel" precipitation) the cumulative infiltration of these soils is described. For both the soils investigated, (grey clay with open cracks and red clays with open cracks), i_I developed in less than 15 minutes.

In discussing the post-infiltration redistribution of water in swelling soils Giraldez (1977) showed hysteretic phenomena, which retard water flow, for water retention and swelling curves. Giraldez (1977) predicted the moisture gradient for a swelling soil is positive. It was shown that swelling soils exhibit a different behaviour from that of nonswelling soils in the redistribution process and consequently runoff generation processes may also be different for swelling soils.

3.5 SHRINK-SWELL PHENOMENA

Smiles and Rosenthal (1968) demonstrated that two bentonite pastes in equilibrium with ambient solutions of the same cation ratio (Na:Ca) but different total cation concentration have different stressstrain relations and therefore different moisture characteristics. The chemical state of a clay as well as its physical state is important in describing the moisture conditions for that clay (see also Grim, 1968). Shrinkage is shown to be linearly related to time and it is suggested that the consolidation (shrinkage) history of material may be necessary to explain the moisture characteristics of a clay. Consolidation has two components: the consequence of interaction of the double layer resulting in elastic behaviour which may be reversible, and spatial reorientations of the colloid particles which probably results in an irreversible change in structure and thus in the properties of the system affecting hydraulic conductivity and diffusivity.

The effect of exchangeable cations on swelling was also discussed by Low and Margheim (1977) who concluded that this effect is secondary to that of surface forces. An equation was developed experimentally which shows that the swelling pressure is an exponential function of water content, the cation exchange capacity and the b dimension of the montmorillonite used in the study. Further analysis showed that the extent of interaction between the water and layer surfaces is independent of the nature of the montmorillonite and that the exchangeable cations make a relatively small contribution to swelling because most of them are not dissociated into the interlayer solution.

A relationship has been derived which shows the connection between the shrinkage curve, the load line, the swelling pressure and the compressibility of clay at any water content (Sposito, 1973, Fig. 3.1). It was found that the shrinkage curve slope ranges from zero to

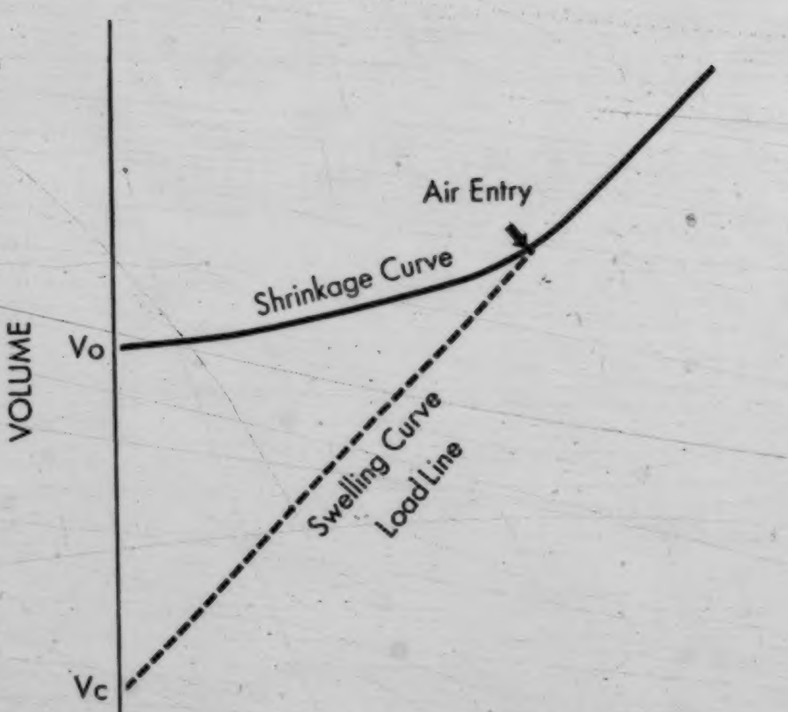


FIG. 3.1 SHRINKAGE CURVE, LOAD LINE AND SWELLING PRESSURE OF CLAY WITH INCREASING WATER CONTENT (from Sposito 1973)

almost one in a monotonic fashion for all swelling clays. The change in slope results from a pressure change as air enters the system. The load line (extrapolation of the shrinkage curve back to the origin) may be interpreted as the swelling curve which corresponds approximately to what would be the reverse of the shrinkage curve under a fixed pressure equal to the swelling pressure (Sposito, 1973).

A theory for the time rate of swelling was suggested by Mesri and Choi (1977). The theory takes into account the increase in swelling index during a load decrement and includes secondary swell during and after the dissipation of excess negative pore pressure. The theory also includes the increase in permeability during a swelling decrement. The theory is illustrated with naturally occurring shales, including that of the Bearpaw formation in Illinois (which also outcrops in the Steveville badlands).

A soil subject to shrink-swell phenomena in a dry state usually has a large number of desiccation cracks. In a saturated state, because of swelling these cracks may become sealed. It has been found that a sodium saturated clay is a much more effective sealer than its calcium counterpart (Dirmeyer and Skinner, 1964). Bridge and Turry (1973) found that gypsum treatments on clays limited the amount of swelling due to the replacement of Na^+ by Ca^{++} on the clay exchange sites. This may lead to a clay being less efficient at sealing desiccation cracks when in the presence of gypsum relative to one which is not. The presence of considerable amounts of gypsum crystals on the surface material in localised areas in the study area may be significant in this context.

Desiccation cracks have been investigated by a number of researchers (e.g. Kimble, 1928; Closs, 1955; Clements, 1957), but by far the most comprehensive experimental study is that of Corte and Higashi (1964). A theory is outlined by Lachenbruch (1960) describing the development of desiccation cracks. Corte and Higashi (1964) examined cracking moisture content, the size of cells made by cracking of the soil, the length of cracks, the number of sides of each cell, and the development of the cracks. They found that crack propogation is very slow compared to that of the fracture of solid materials due to the desiccation of the soil. Cracks propogate more slowly in thick soil than in thin soil horizons. The crack area in the field is an indicator of the shrinkage of the soil under the constraints of bottom and marginal adhesion. Actual shrinkage is always smaller than free shrinkage (shrinkage as usually measured in the laboratory) which Corte and Higashi (1964) interpret as elastic stress existing in the soil which enables the crack generation and development to continue. Apparently repetition of desiccation and wetting of the soil does not make any difference in the final features of soil cracking. It seems cracks start slightly below the surface and the moisture profile itself has no definite influence on cracking phenomena. Corte and Higashi (1964) noted that some cracks appeared to begin from air bubbles. Complete cover of the soil surface by layers of shale results in "habituation", i.e. repetition of crack patterns when desiccation and wetting are repeated. Mixture of gravel into the soil in the cracks creates strong adhesion of the soil, and crack patterns avoid the former traces of cracks when the soil is desiccated again. It was concluded that hexagonal patterns predominate only in very thin (less than 4 mm) layers of soil, the more common pattern being polygonal.

It should be stressed that most of the work on shrink-swell phenomena has been conducted in the laboratory under extremely precise, controlled conditions. Little effort seems to have been made to translate the results of such experiments into the field. The question of the applicability of these results in the field cannot be overemphasised. Read, for example, (cited in Collis-George, 1977) found that unstable swelling clay did not undergo free vertical expansion in a laboratory column as it would in the field, because of the mechanical interaction of the material with the walls of the container. This apparently leads to a zone of low hydraulic conductivity, the "infiltration throttle," being produced (Collis-George, 1977). The effects of soil cracking were investigated by Marshall (1959) who found that if the soil is heavily cracked then infiltration rates may be very high. Cracks were often not completely closed even months after wetting. Laboratory measurements indicated low hydraulic conductivity, while in the field, because of cracking, it was suggested that conductivities may be considerably higher. Studies of heavy textured soils have shown that field permeability exceeds laboratory permeability by several orders of magnitude (Bear, et al., 1968).

3.6 BADLANDS RESEARCH

Artificial runoff has been studied in a number of semiarid and badland areas. Sprinklers have been used both in the laboratory and the field to simulate rainfall as well as more primitive methods such as simply pouring water from a can onto the surface to observe what happens. Few people have investigated naturally occurring runoff processes. This may be partly due to the intermittent nature of events, but may also be due to the complexities involved.

Emmett (1970) carried out a comprehensive study of the hydraulics of overland flow both in the laboratory and in the field using sprinklers in a semiarid area. He noted that surface runoff tended to concentrate in several lateral concentrations; however, these concentrations were not considered to be rill flow.

> "The general appearance of runoff at most field sites was one of omnipresent surface detention, easily detected by the glistening of a sheet of water in the sunlight."
>
> (Emmett, 1970, p. A26)

Emmett (1970) attempted to observe rill formation with his sprinkling tests and found that after almost 10 hours of sprinkling at an intensity of 8.5 ins/hr (216 mm/hr) no observable rills had been formed on poorly sorted sediments ranging from clay size to small gravel with sagebrush cover removed to ground surface level. Even when the intensity was raised to 10.5 ins/hr (267 mm/hr) and continued for six hours no rills appeared. It appeared that rills may or may not develop on unrilled surfaces if some threshold is exceeded which causes a change in the erosion rate. It was noted, however, that after 10 years of research in the area and

> "despite efforts to observe overland flow from thunderstorms occurring during the several weeks at the project

> area each year, overland flow was never observed in the field."
>
> (Emmett, 1970 p.A36)

In a series of papers Schumm (1956a, 1956b, 1964) and Schumm and Lusby (1963) discuss the seasonality of infiltration rates on hillslopes. Through investigating the infiltration rates of the Chadron (comparable to the "popcorned" surfaces in the study area) and Brule (comparable to the clay-cemented sandstones in the study area) formations, Schumm (1956a, 1956b) concluded that topographic differences could be explained by the differences in infiltration rates of the slope surfaces. A hand pump was used to spray water on the slopes. On the Brule formation, surface runoff occurred almost immediately via rill networks. In contrast, 4.5 gallons were sprayed onto a similar area (six square feet) of the Chadron surface before surface runoff began. Water disappeared rapidly between the aggregates and flowed in subsurface channels to reappear at the base of the slope and continue as surface flow across the pediment. It was concluded that erosion of the Brule slopes was mainly by rainwash whereas that of the Chadron was by creep. It was suggested that slope reduction occurs twice as rapidly by rainwash as by creep. Basal sapping at the base of the slopes resulted in pediment formation.

> "This concept may be of importance in attempting to extend the relationships between geomorphic and hydrologic characteristics of drainage basins in semiarid regions to more humid areas and also as a partial explanation of the high rates of erosion in semiarid regions."
>
> (Schumm, 1956a p.704)

In a wider context it was suggested that areas where creep dominates over rainwash, and vice versa, may be end members of a continuous series ranging through all proportions of the two processes depending on prevailing vegetation, soils and climate. It was concluded that as the time required to initiate runoff increases the slope angle decreases, or alternatively, as infiltration rates increase slope angles decrease. Apparently "infiltration rates are one of the most important factors influencing topographic development" (Schumm, 1956b, p.645). Schumm further stated (1956a) that five percent silt in the material causes enough capillary cohesion in the soil to maintain steep slope angles, and the permeability is low enough (he measured 0.004 cm/sec) to aid surface runoff. Capillarity was reported to have an approximate rise of about five feet, in his study area. The Chadron formation, with its layer of loose aggregates ("popcorn") was compared to profiles in humid areas, the "popcorn" layer forming a mulch which is stated to be analagous to vegetation and deep soil profiles of humid areas.

It was stated (Schumm, 1956b) that snow plays an important role in the destruction of the relatively impermeable surface formed during the summer by rainbeat. Melting of snow wets the soil, allowing frost action to occur. It was stated, however, that most of the snow cover is lost by sublimation and therefore water content of the soil is low (Schumm, 1964). Although not mentioned, this would surely limit the amount of frost action occurring in the soil. Several periods of freezing and thawing apparently change the less permeable rilled surface to a highly permeable surface composed of soil aggregates without rills. Swelling and shrinking of the soil was considered to be responsible for the creep measured on the slopes during the summer. Infiltration rates measured with a Rocky Mountain infiltrometer increased from 0.89 to 1.12 ins/hr (23 to 28 mm/hr) for dry runs and 0.67 to 0.89 ins/hr (18 to 23 mm/hr) for wet runs from 1953,54 to 1958 increasing from 0.19 to 0.35 inches (5 to 9 mm). Only about one third of the rainfall that fell in 1953,54 summer, however, fell in the summer of 1958. The differences were attributed to the smaller amount of rainfall in 1958 causing less compaction and therefore allowing higher rates of infiltration. However, in spite of the compaction and surface sealing after 1.94 inches (50 mm) of rain, the surface absorbed 0.16 inches (4 mm) more water before runoff began in 1958 than in 1953,54 and the infiltration rate was 0.23 ins/hr (6 mm/hr) greater in the ungrazed catchments. It was concluded that antecedant moisture effects were negligible. It was stated that the dominant factor causing the seasonality of soil behaviour was the lack of vegetation cover. Seasonal trends were also found in an erosion study on the Steveville badlands, (Campbell, 1970a) confirming to some extent the findings of Schumm (1964).

A number of papers have been concerned with the development of piping in badland areas. Several conclusions have been reached which relate to runoff processes and water movement generally in these areas. For piping to occur several factors seem necessary: the alternation of consolidated and resistant beds with softer less resistant beds (Cockfield and Buckham, 1946; Whipkey, 1965); the presence of montmorillonite clay (Mears, 1963; Parker, 1963; Ward, 1966; Heede, 1971); cracks in the soil and jointing and bedding (Buckham and Cockfield, 1950; Mears, 1963; Parker, 1963; Heede, 1971); aperiodic, high intensity rainfall (Parker, 1963; Ward, 1966; Heede, 1971). Apparently there is no agreement on what causes the initial concentration of water in the soil or surficial geologic material in order to initiate pipe development. The concentration at various levels has been attributed to an impermeable layer beneath an easily eroded layer (Fletcher, 1954; Ward, 1966); cracking and jointing or bedding (Buckham and Cockfield, 1950; Parker, 1963); and the height of the seasonal water table (Rubey, 1928; Buckham and Cockfield, 1950).

In discussing clays in the Milk River Canyon, Alberta, Barendregt and Ongley (1977) suggest that throughflow is possible and important in the initiation of piping. Throughflow was suggested to occur through a process of eluviation with the coarser grains providing a skeleton matrix through which the finer grains and water may move. The presence of impermeable layers in the clays was discussed but it was concluded that the water and frost action would eventually open pathways through the layers causing the head of water above it to drain suddenly and thus destroy the effectiveness of the layers permeability. Although they considered their study area to be unique in Alberta, suggesting that extensive piping did not occur elsewhere, this is definitely not the case.

A series of sprinkling tests were carried out on various lithologies of the present study area (Bryan, et al., 1978). In all tests conducted the properties of the surface materials rather than the rainfall controlled the patterns of surface and subsurface flow. Three different erosion processes were recognised: surface erosion; subsurface erosion; and mudflows. Threshold amounts of rainfall for the initiation of runoff on an initially dry surface appeared to be about three to five millimetres on shale units with only about half a millimetre on sandstone and micro-pediments. About half this amount of rainfall was required for wet surfaces. Since the runoff generation on some plots was non-uniform, the concept of partial area contributions to runoff was invoked. Surface moisture contents showed little variation and depth of wetting did not exceed three millimetres on any of the lithologies. Variations in response to rainfall were therefore attributed to differences in structure and composition of the surface material rather than variations in soil moisture conditions prior to the onset of rainfall. The quick response of rills to rainfall was attributed to rapid sealing of shallow cracks combined with the depositional layers of compacted clay which limited further infiltration. Bryan, et al., (1978) concluded that runoff generation, and to some extent rate of runoff are controlled by depositional processes as "these are the only processes capable of creating the surface structures underlying the rills which inhibit infiltration and favour runoff generation" (Bryan, et al., 1978). Since rillflow and pipeflow are closely related, their individual peaks cannot be identified on storm hydrographs and it was concluded that it is therefore difficult to assess their relative contributions to runoff.

It is believed that only ten percent of the rain that falls in the semiarid zone is sufficient to produce runoff (Slatyer and Mabbutt, 1965). Peel (1975) suggested that

> "if the rapid generation of desert floods cannot be attributed to anything exceptional about desert rainfall, the answer must lie in the characteristics of the terrain."
> (Peel, 1975 p.121)

He discussed the low infiltration capacities of the desert surfaces and runoff thresholds and concludes that the physical composition of the upper portion of the regolith is the main factor in arid land runoff generation. It was suggested that low runoff thresholds are evident in semiarid and arid areas, although this has yet to be proved, and constitutes a major part of the present investigation.

From previous work conducted in the badlands it is known that runoff occurs in the channel after some threshold amount of rainfall has been achieved. Generally the amount of rainfall occurring before runoff appears in the channel is approximately three to four millimetres (Campbell, 1978). Bryan, _et al._, (1978) found similar amounts of rain from sprinkling experiments were required to produce runoff on small plots, the amount varying with the lithology.

3.7 MOISTURE CONTENT - RAINFALL RELATIONSHIPS

Traditionally, various soil types are regarded as having characteristic moisture retention capacities. In a simplistic view of the runoff and infiltration processes the moisture content should be well correlated with the amount of rainfall from a given storm. In other words, for a given soil type one would expect to see increasing moisture content with increasing rainfall amount. Moisture content of the soil would increase up to a point at which the soil becomes saturated and no more water can be taken up by the soil. This threshold level, equivalent to the water capacity of the soil, may be interpreted as the point at which surface runoff begins, in terms of saturated overland flow model (Kirkby and Chorley, 1967).

Grim (1968) presented a series of curves for the different clays of water sorption percent versus time. These curves are essentially straight for the first one and one half minutes of sorption when 60, 70, and 90 percent sorption has occurred for illite, calcium kaolinite and sodium kaolinite respectively. After this time there is a sharp break and the curves tend to flatten out. The point at which the curves flatten is approximated to the liquid limit of the clays. Calcium montmorillonite shows similar characteristics, with the curve flattening at about 170 percent water sorption. With sodium montmorillonite, however, water sorption continues well beyond the liquid limit. Grim (1968) suggested the liquid limit may be approximated to the maximum water holding capacity since calcium occurs in most clays and the break in the water sorption curve is particularly sharp for calcium saturated clays. One would expect, therefore, that the moisture content of soils high in clay content, would show a similar pattern in that the curve of moisture content versus rainfall amount should rise approximately as a straight line relation and then flatten out around a threshold rainfall amount as the liquid limit or maximum water capacity of the soil is reached.

According to Horton's (1945) theory of infiltration and runoff, the moisture content should be highly correlated with rainfall intensity, with moisture content increasing with intensity until intensity exceeds the infiltration rate at which point the straight line relation would cease and the curve would flatten markedly. The duration of rainfall should also be well correlated with moisture content, with the moisture content increasing as storm duration increases. In terms of the saturation-overland flow model this relation would hold until the soil became saturated at which point runoff would occur and the curve would flatten out.

A correlation was found between rainfall amount, intensity, duration, and antecedant soil moisture and infiltration (Sagi, 1969). Canarche, *et al.*, (1969) found correlations between infiltration rate and percent of water free pores at field capacity, as well as between infiltration rate and log hydraulic conductivity. No correlations were found between infiltration rate and antecedant moisture content. Sagi (1969) warns against the comparison of results from natural rainfall events with those from sprinkler irrigation.

Vûcíc (1969) found a definite correlation between the mechanical composition of a soil and the infiltration rate for non-structured or insufficiently structured soils. "It is also known that soil structure changes the relation to a great extent and increases the infiltration rate" (p. 344). In soils with the same mechanical and structural composition he suggested that the infiltration rate is influenced by the stability of the aggregates. A number of other studies have related soil texture to moisture characteristics and permeability (e.g. Aronovici, 1946; Salter and Williams, 1965, 1967). The prediction of permeability from such physical properties of the soil has been approached by relating permeability and pore size distribution (Hillel, 1971). The results of these theories[2], however,

> "while more generally applicable than those based on earlier models, still appear to be valid only for certain coarse materials in which capillary phenomena predominate."
>
> (Hillel, 1971 p.98)

Footnotes:

[1] Hysteresis may be defined as the phenomena whereby "... the state of the medium depends on its previous history as well as the instantaneous value of the applied force" (American Geological Institute, 1974 p.245). In the case of soil water, the applied force is matric suction. For more details, see Hillel (1971) and Childs (1969).

[2] See Childs and Collis-George (1950), Marshall (1958) and Millington and Quirk (1959) for more details.

CHAPTER IV

METHODS

4.1 INTRODUCTION

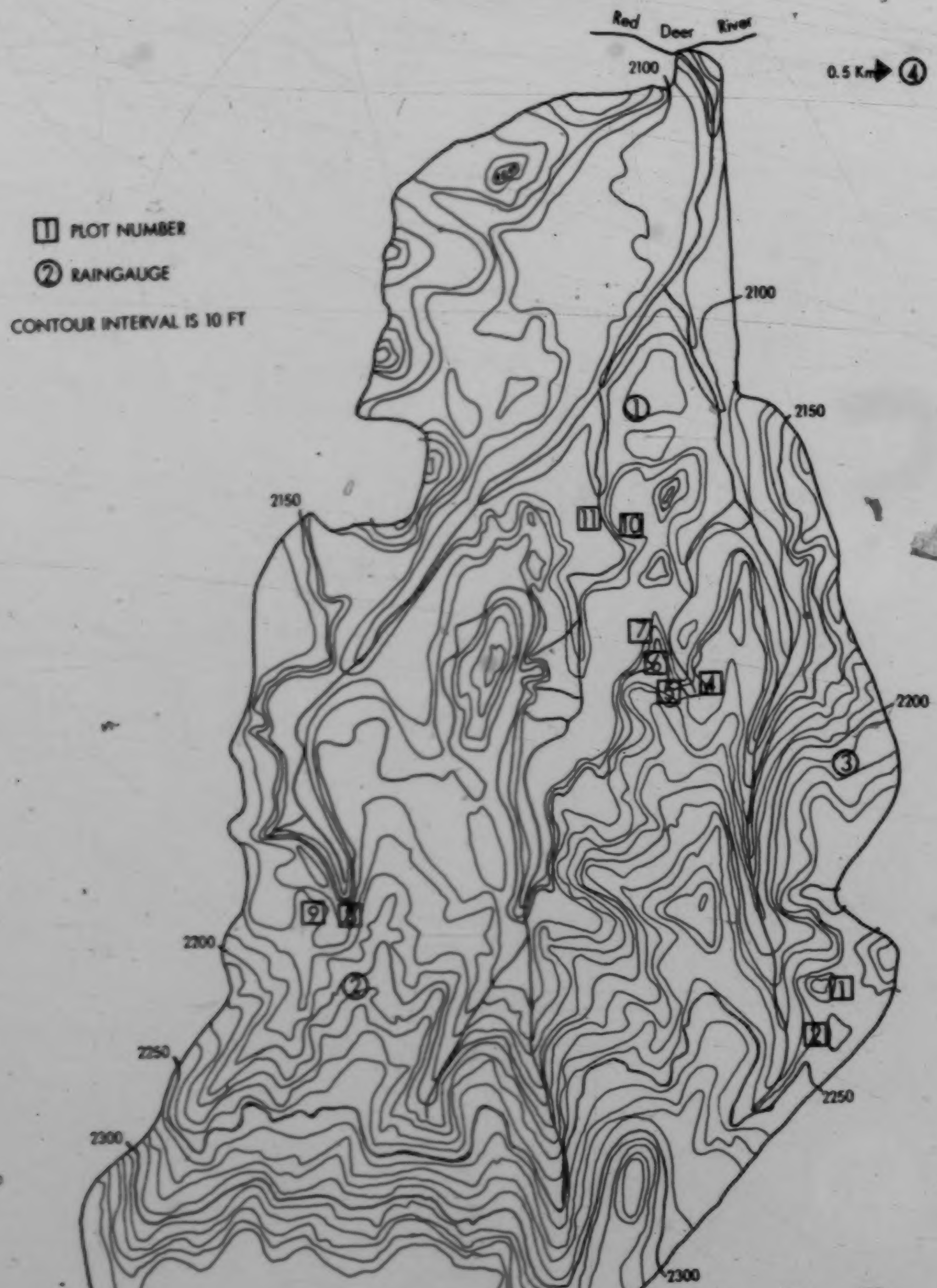
Both field and laboratory investigations were carried out in the course of this study. The laboratory work was intended to supplement field information by attempting to replicate some field conditions in a controlled state. The nature of the material, however, makes reproduction of the field situation extremely difficult since it was not possible to collect "undisturbed" samples. This is due to the extremely delicate and complex physical structure of the material. For example, structural relationships between desiccation cracks in the upper "popcorned" surface and intermediate crustal layer on the shale lithology was impossible to duplicate in the laboratory. In general then, the laboratory tests provide information only for comparative purposes and lend support to some qualitative observations. A discussion of the methods of analysis follows the description of field and laboratory methods.

4.2 FIELD MAPPING

Selection of sampling sites for the investigation of the various terrain types required the preparation of maps of the study basin showing the geology and lithology and the extent of the vegetation cover. Information collected during mapping was used for selection of the plots from which further sampling was carried out.

A ten foot contour interval topographic map (Fig. 4.1) was produced

FIG. 4.1 CONTOUR MAP of Study Basin

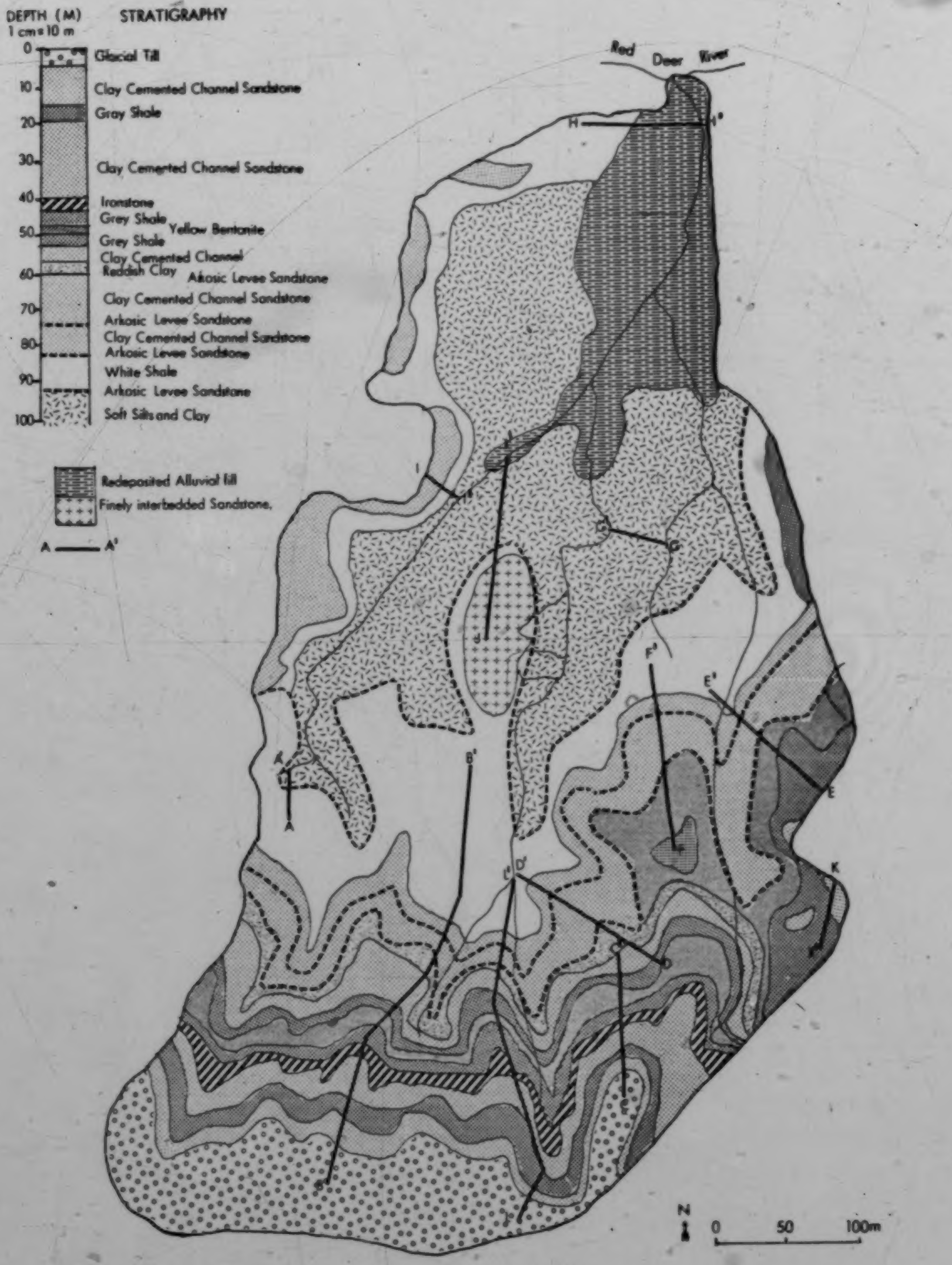

Red Deer River
2100
0.5 Km → ④
1 PLOT NUMBER
② RAINGAUGE
CONTOUR INTERVAL IS 10 FT
2100
①
2150
2150
11
10
7
6
5
4
2200
③
9
8
2200
②
1
2
2250
2250
2300
2300
N
0
50
100m

based on a contour map (20 foot contour interval) prepared by the Department of Lands and Forests (Aerial Surveys Section) as project 36/61, January 1964. This map was based on aerial photography taken in 1961. The Lands and Forests map was at a scale of one inch to 400 feet, but for a base map of the study basin, the scale was enlarged to one inch to 200 feet. Additional contours and refinements of the Lands and Forests map were made following field surveys.

A geology and lithology map (Fig. 4.2) of the basin was prepared from twelve stratigraphic sections mapped within the basin. Geologic units were defined in the field largely on the basis of colour and changes in slope angle as well as lithologic differences. The same scale was applied to mapping the surficial geology as that used for defining lithologic units on the various plots used for sampling. Slope angles were measured with an abney level and graduated stakes with divisions every 10 cm. A line level was used for levelling when required.

Since the strata are almost horizontal (Dodson, 1970), once the sections had been mapped a surficial outcrop map (Fig. 4.2) could be drawn up by extrapolation along the contours (Fig. 4.1). The twelve stratigraphic sections are located on Fig. 4.2 (A - A', ..., L - L'). Several distinct units were evident in mapping. In all cases where the clay cemented sandstone outcrops it is interbedded with clay ironstone bands. The uppermost clay cemented sandstone could be traced by eye across the basin. This bed, beneath about three metres of glacial till, averages about ten metres thick. Beneath this is a seven metre

FIG. 4.2 SURFICIAL GEOLOGY AND STRATIGRAPHY


DEPTH (M)
1 cm = 10 m
STRATIGRAPHY
0
10
20
30
40
50
60
70
80
90
100
Glacial Till
Clay Cemented Channel Sandstone
Gray Shale
Clay Cemented Channel Sandstone
Ironstone
Grey Shale
Yellow Bentonite
Grey Shale
Clay Cemented Channel
Reddish Clay
Arkosic Levee Sandstone
Clay Cemented Channel Sandstone
Arkosic Levee Sandstone
Clay Cemented Channel Sandstone
Arkosic Levee Sandstone
White Shale
Arkosic Levee Sandstone
Soft Silts and Clay
Redeposited Alluvial fill
Finely interbedded Sandstone.
A — A'
Red Deer River
H
H'
I
I'
G'
G
F'
E'
E
A'
A
B'
K
C'
D'
N
0
50
100m

band of grey shale overlying a twenty metre deposit of clay cemented sandstone. Below this is a five metre band of clay cemented sandstone with rather more ironstone interbedded than other sandstone beds and is therefore designated as an ironstone band interbedded with sandstone (Fig. 4.2). Below this are two, four metre bands of grey shale between which is a two metre band of yellow bentonite.

Underlying the grey shale is a massive outcrop, thirty metres thick, of clay cemented sandstone interbedded with ironstone layers. A band of reddish clay (four metres thick) appears in the upper section of this which is bound on its base by a thin band of arkosic sandstone, continuous across the basin. Fifteen metres below this is a second continuous band of arkosic sandstone.

Beneath this major unit is an outcrop of white shale generally ten metres thick but quite variable in thickness across the basin. Underlying this unit, are soft, yellowish silts and clays forming the last structural unit in the basin. Alluvial fill (fine silt and clay) has been deposited in the lower channel area of the basin. A large residual mound in the middle of the basin has a complex outcrop geology of finely interbedded clay cemented sandstone, ironstone and clays. This emphasises the complex nature of bedding across the basin in general as well as in the entire badlands area (Faulkner, 1970).

Generally the major beds are continuous, although some pinching out of beds occurs throughout the basin (Fig. 4.3). Frequently the beds also vary considerably in thickness over the basin which further complicates detailed mapping.

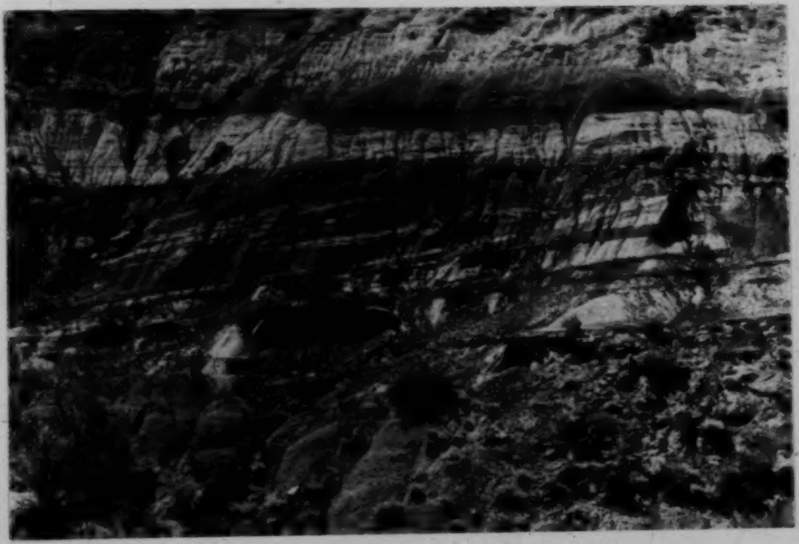


Fig. 4.3. Pinching out of strata of the Oldman formation in the study basin. The dark thin bands are layers of ironstone.

Aerial photographs (1969, 1":1,000') were used in the preparation of a vegetation map (Fig. 4.4). This was supplemented with field surveying. Percent area cover was estimated from field investigation using the chart for estimating percentage of mottles in soils published with Munsell's (1967) standard soil colour chart.

In general, vegetation coverage is fairly high compared to that normally expected for badlands (e.g. Schumm, 1956a; Weyman, 1975). It is believed, however, that this basin is representative of the larger badlands area of the Red Deer River valley and of other areas (Campbell, 1978). The reason badlands are generally considered to be devoid of vegetation may be because bare slopes appear more spectacular than more "ordinary" vegetated slopes and therefore appear more dominant to the observer.

Forty two percent of the basin area has a 50-100 percent vegetation cover. Nine percent of the basin area has a 25-50 percent cover; three percent has a 10-25 percent cover; seven percent has a 1-10 percent cover; and 23 percent has 0-1 percent cover. Only 18 percent of the basin area is devoid of vegetation.

In this highly sensitive environment vegetation is an important control on erosion rates and also affects surface flow by increasing infiltration (Blackburn, 1973) and offering obstruction to flow paths. In many cases it serves as a protective cap over soft material (Fig. 4.5).

4.3 TERRAIN TYPES

The combination of lithology, geomorphology and vegetation in the

FIG. 4.4 VEGETATION Map of Study Basin

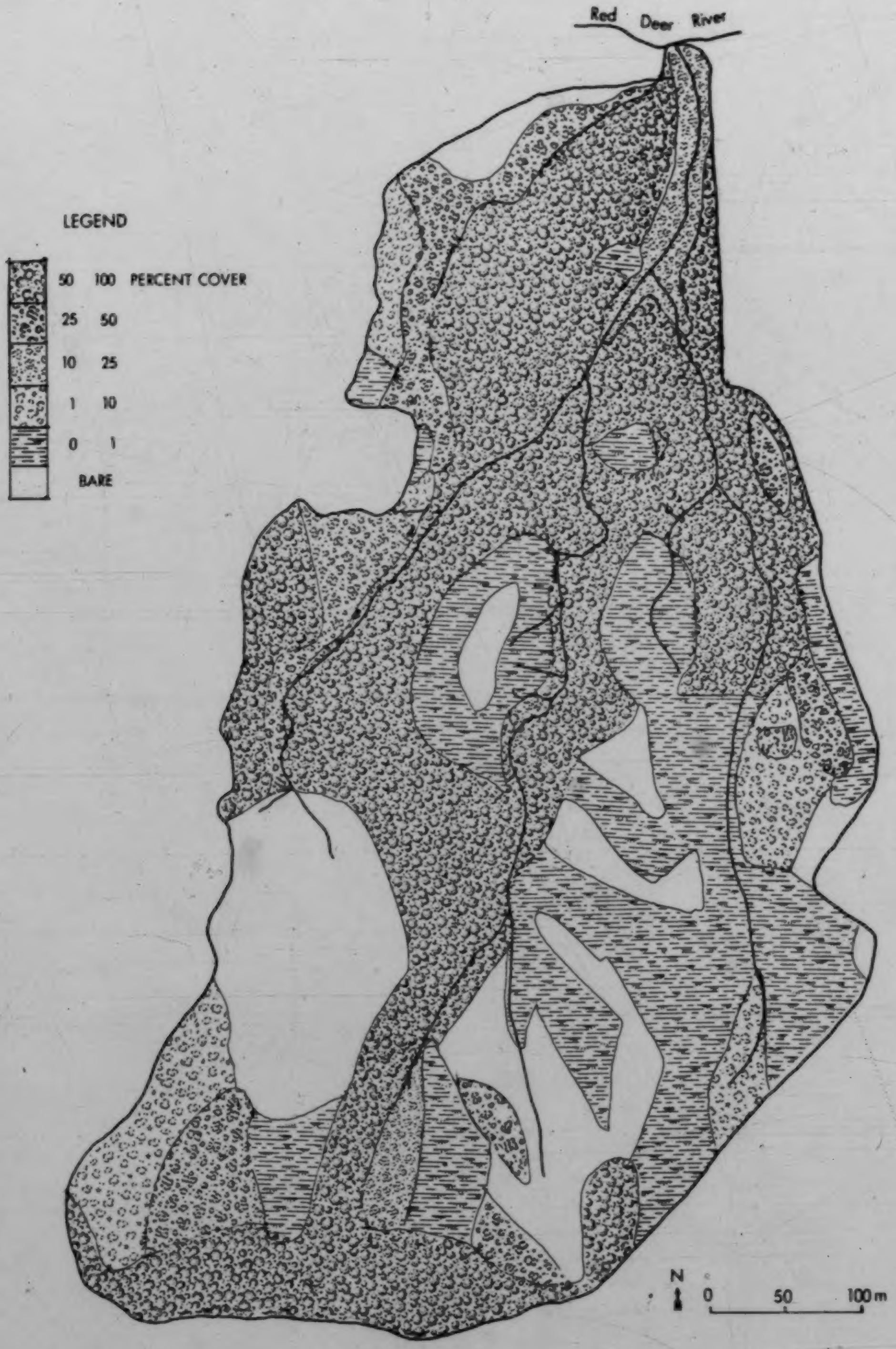

Red Deer River
LEGEND
50 100 PERCENT COVER
25 50
10 25
1 10
0 1
BARE
N
0 50 100 m

Fig. 4.5a. Vegetation (greasewood) serving as a protective cap over softer material (shale) beneath.

Fig. 4.5b. Vegetation (lichen) serving as a protective cap over softer material (micro-pediment) beneath.

study basin defines four major terrain types. These are:

1. The desiccated "popcorn" surface of the shale slopes (Fig. 2.6).
2. The highly rilled sandstone slopes (Fig. 2.4).
3. The micro-pediment surfaces (Fig. 2.10).
4. The soils with a vegetation cover (Fig. 2.11).

More so than the other terrain types the "popcorn" surface covers a range of lithologies, particularly with regard to clay mineralogy. These are well recognised in the field by abrupt colour changes of the material as well as through differing desiccation crack patterns and densities.

These terrain types were the units selected for comparative study of their near surface hydrologic properties. Sampling techniques were adopted to include all these surfaces and to include the range of lithologies within a terrain type.

4.4 SELECTION OF PLOTS

To observe the effects of aspect, a number of paired plots with each member of the pair having the opposite aspect to its partner, were selected within the basin (Fig. 4.1). These plots covered the range of terrain types and lithologies encountered within the basin. Each plot of the pair had similar slopes, but slopes varied between pairs. Since the basin main axis was aligned north-south, resulting in the topographic grain running north-south, and the prevailing wind is from the southwest, plots were selected with mainly east and west facing aspects. Beaty (1975b) stresses the importance of wind in distribution of rainfall and snow and relates this to coulee alignment in the area (dominantly southwest-northeast). Apparently wind dominates other variables (e.g. solar radiation) relating to aspect. Consequently, east and west aspects should show the most evidence of "aspect effect" if any such effect exists. In most cases a single plot covered a range of lithologies easily identified by colour changes in the material and/or breaks in slope. Thus, the variables of lithology, aspect and slope within and between plots could all be taken into account for comparison of their near surface hydrologic properties. Details of individual plots are set out in Appendix I.

Each plot was one metre wide and the length of slope varied according to the general form of the surface, but ranged from 1.2 m to 4.1 m. Accessibility for sampling was a major factor in locating each plot and defining the length of slope to be sampled. Since much of the basin consists of highly bentonitic materials almost all surfaces in the area were extremely slippery when wet and care had to be taken when selecting sampling sites to ensure reasonable access in wet conditions. Ten plots were established with a total of 38 sampling sites. A possible 18 different lithologies were recognised and used in further laboratory tests.

4.5 PRECIPITATION MEASUREMENT

Three 12 inch (30.5 cm) Belfort recording rain gauges were set up in the study basin with a fourth (13 cm) Lambrecht recording gauge located at the Park headquarters (Fig. 4.1). One Belfort (gauge No. 1) and the Lambrecht (gauge No. 4) were on weekly charts while the other two gauges were on monthly charts. The two weekly charts provided detailed information on rainfall intensities and duration for the study period.

4.6 FIELD MOISTURE SAMPLING

Moisture samples were obtained from each plot prior to a rainstorm (antecedant) and as soon after cessation of rainfall as possible (post-rainfall). Samples were taken from each lithologic unit on each plot at both the surface (1 - 5 cm depth) and subsurface (5 - 10 cm depth) levels. Sample locations were chosen at each sampling, but since the extraction of samples destroyed that particular location on the plot, these were not strictly random. As the study progressed it became obvious that almost in every case the soil was "dry" (i.e. at an extremely low water potential) before each period of rainfall. To save time in the field and to obtain results from a greater number of rainfall events (due to the suddeness of storms antecedant moisture sampling was not undertaken as it was not always obvious when rain was about to occur) a number of antecedant moisture contents were deemed "zero" at the outset.

Samples of approximately 150 g were taken, sealed in plastic bags and labelled. These were weighed wherever possible in the field on a beam balance or, in the laboratory on a Mettler top loading balance. All measurements were read to within 0.01 g. All samples that had to be transported to the laboratory before weighing were double-bagged and double sealed to ensure moisture retention. The weight of the samples minus the weight of the bags was deemed the "wet weight" of the sample used for later analysis. Depth of wetting, taken to be the point of maximum advance of the wetting front, was also measured when and where the moisture sample was taken.

4.7 PLOT PROFILES

In order to obtain a record of the form of each sampling plot detailed measurements of slope angles and relative relief were made. Slope angles were measured over one metre lengths on both north and south sides of each plot using an Abney level. Relative relief was measured using a levelled one metre stake across the profile with vertical (90°) measurements to the surface recorded every 10 cm across the plot. This was repeated every 10 cm up the length of the plot also. Detailed plot profiles appear in Appendix I.

4.8 CRACK PATTERNS

Random grid coordinates of a one square metre grid were selected using a random number table, along which were measured cell (a cell is defined as that portion of material bounded on all sides by a desiccation crack) size and width of desiccation cracks to within one millimetre. These were then measured every 10 cm along the randomly selected grid line for each lithology on the plot. Cell size was obtained by measuring the longest and shortest axes and using the product as the size of the cell. In all cases, measurements were taken at dry moisture conditions. Details of the crack patterns for each lithology also appear in Appendix I.

4.9 RILL COUNTS

Quantitative evidence for surface erosion (and hence overland flow) as opposed to subsurface erosion (indicating subsurface throughflow and/

or pipe flow), was obtained by using rill frequency as an index. On two broadly defined lithologic units of shales and clay cemented sandstones the number of rills were counted along two metre widths of slope units where a sand unit occurred above a clay unit and vice versa. Counts were made randomly within the study basin. Only ten counts were made since the dominance of rills on the sandstone was clearly obvious.

4.10 MOISTURE DETERMINATION

Determination of water content was carried out using the gravimetric method according to Gardner (1965):

1. Having obtained wet weights of the samples (4.6) taken in the field, the samples were placed in an oven at 105°C for 24 hours.
2. Samples were removed and weighed immediately to determine the dry weight.
3. Water content of the sample on a dry weight basis (θdw) was then computed according to

$$\theta dw \quad \frac{\text{weight of wet soil}}{\text{weight of dry soil}} - 1$$

By multiplying θdw by 100 the percentage moisture content was obtained.

The same method was used for all samples so that results could be directly compared (Gardner, 1965). Although this method necessarily destroys the sampling site and it is time consuming and laborious, it is reproducible and simple to use as well as being the "traditional" method employed for obtaining the soil moisture content (Hillel, 1971).

The moisture content contributed to the soil from each rainfall was computed according to

$$\text{Post } \theta dw - \text{Ant } \theta dw = \quad \theta dw$$

where Post θdw is the moisture content measured in samples taken immediately following a rainstorm; Ant θdw is the moisture content measured in samples taken immediately prior to a rainstorm and θdw is the change in moisture content as a consequence of the rainstorm.

4.11 SPECIFIC SURFACE

Specific surface refers to area per unit weight of material. A number of methods of measurement are described by Mortland and Kemper (1965). A common method is via the absorption of a monomolecular layer of ethylene glycol. A relatively rapid method using ethylene glycol monoethyl ether (EGME) was introduced by Carter, et al., (1965) and adopted for use with soils by Heilman, et al., (1965). Since the latters' method is relatively simple and rapid this method was chosen for measurement of the specific surface of the "soils" in the study basin. The following procedure was adopted:

1. Five gram samples of "soil" were ground to pass a 60 mesh sieve.
2. The sample was then treated with 10 ml of 30 percent hydrogen peroxide and 20 ml of water. After about 15 minutes, when reaction had ceased, the samples were calcium saturated with N $CaCl_2$ using a Buchner funnel and suction. After saturation the sample was washed three times with distilled water to remove excess salt. The sample was then oven dried and again ground to pass a 60 mesh sieve.
3. After pretreatment, the samples were dried to constant weight over

P_2O_5 in an evacuated desiccator.

4. Duplicate one gram samples were weighed out and placed in small aluminium pans. The samples were treated with EGME to form a soil/EGME slurry that was allowed to equilibrate for 30 minutes in the desiccator over $CaCl_2$. The desiccator was then evacuated repeatedly until constant weight of the samples was reached (three evacuations were required).
5. Grams of absorbate retained per gram of soil was obtained and the specific surface in m^2/g calculated by dividing by 0.000286 g/m^2 (Heilman, et al., 1965).

4.12 PARTICLE SIZE ANALYSIS

Dispersion of soil aggregates and the formation of a stable suspension is required in order to quantitatively separate the clay fraction from the bulk soil sample. Genrich and Bremmer (1974) describe the dispersion of soils without use of chemicals for removal of organic matter, Fe, Al, and Si sesquioxides and $CaCO_3$, which essentially involves ultrasonic vibration. This method avoids the time consuming chemical pretreatments necessary and allows almost quantitative recovery of soil material in the various fractions isolated.

1. Approximately 60 g of air dried soil was allowed to soak overnight in distilled water.
2. The samples were transferred to 600 ml beakers, with 250 ml of distilled water and treated with ultrasonic vibration using the probe-type vibrator (Genrich and Bremmer, 1972) for approximately five minutes.

3. The sonified sample was transferred to a 2,000 ml beaker and distilled water added to the 1,500 ml level.
4. The clay (less than 2 µm) fraction was separated by gravity sedimentation (Jackson, 1956) using only distilled water, Extraction depths and time were calculated according to Tanner and Jackson (1947). About six extractions were necessary to remove all the clay.
5. After all the clay was extracted, the residue was wet sieved using a 230 mesh sieve to separate the silt and sand. The silt and sand fractions were oven dried, weighed and percent sand, percent silt and percent clay calculated.

4.13 CLAY MINERALOGY

After the clay fraction was separated in the particle size analysis procedure, it was Ca-saturated with N $CaCl_2$ to flocculate the suspension. From the total flocculated clay fraction for a given soil, a depth integrated sample was taken to be used for mineralogical analysis. This sample was centrifuged with distilled water three times to remove excess salts and slides prepared for X-ray diffraction according to the "paste method" (McKeague, 1976). X-ray diffraction of the clay samples was done by the Alberta Research Council.

4.14 AGGREGATE STABILITY

Because of the wide range of disintegrating forces the choice of a particular method of determining aggregate stability depends on the interpretation to which the results are put (Baver, et al., 1972;

Kemper, 1965). Since raindrop impact is likely to be important in furthering the disintegration of surface aggregates in the badland "soils", a method involving simulated raindrops may be preferable (Kemper, 1965). Undirectional wetting and differential hydration not important in the commonly used flood wetting technique may be significant in montmorillonite-rich soils, such as the "soils" of the badlands and therefore a drop-testing technique, being a closer simulation of natural processes may improve predictive capacity of soil aggregation (Bryan, 1976). Such a technique was developed by McCalla (1944) and adopted for use here:

1. A soil aggregate of about 0.15 g was placed on a one millimetre sieve and drops of distilled water falling 30 cm from a burette were allowed to strike it.
2. When the aggregate was at the point of being washed through the sieve, it was considered destroyed.
3. The number of drops per 0.1 g of material required to destroy the aggregate was calculated. This test was repeated 20 times for each soil type investigated.

In soils with high exchangeable sodium oven drying may cause a transient stability of otherwise unstable aggregates. This transient stability, however, decreases with increasing time after oven drying (Kemper, 1965). Although the soils used in this study were oven dried and possibly contain large amounts of exchangeable sodium, it was at least one month for all "soils" since the oven drying took place before this test was conducted, and therefore this was not considered to be a problem. Since this technique is very laborious and time consuming, and in view of the other laboratory results which did not prove very useful, a complete analysis of all 18 lithologies was not attempted and only tentative measurements of aggregate stability were obtained (see Chapter V).

4.15 SHRINKAGE

Determination of percent shrinkage was carried out according to A.S.T.M. (1970):

1. Samples were passed through a 60 mesh sieve and wetted thoroughly to their approximate liquid limits.
2. Shrinkage dishes were greased on the inside with vaseline to avoid adhesion of the "soil" and filled with the wetted sample. Air in the sample was expelled by tapping the dish on the bench and then the sample was levelled off with a straight edge. Samples were immediately weighed and left to air dry after which they were oven dried and weighed again.
3. Volume of the shrinkage dish and the dried soil cake was obtained by mercury displacement.
4. Percent shrinkage was calculated based on the original volume of the wet soil.
5. Approximate Atterberg liquid limits were obtained by calculation of the gravimetric moisture content (4.3.1) of the soil cake on a dry weight basis.

4.16 METHODS OF ANALYSIS

All data analyses were carried out using the Midas Statistical

Research Laboratory. Initially data were generally described to obtain simple statistics such as means and standard deviations. Further analysis was then performed.

In any statistical analysis of data, the first task required is to test for the normality of the data in order to check if later assumptions made in parametric tests may be valid or not. This was done for all data using the skewness and kurtosis tests (Snedecor and Cochran, 1967).

To statistically test relationships between moisture content and rainfall amount, duration and intensity, simple and multiple linear regressions, calculated by the least squares method were used. Various groupings of the data were tried in an attempt to improve estimates.

In view of the fact that non-normality of the data, discussed previously, may possibly affect the results obtained from the regression model, non-parametric tests were also used to investigate the relation between sampling sites in order to group data, as well as the relation between moisture content of each sampling site and rainfall amounts. By ranking data it was thought that sites giving consistently high or consistently low moisture values may also be detected. Spearman's rank correlation test (Hammond and McCullagh, 1974) was carried out on each sampling site between moisture content and rainfall amount. The Kruskal-Wallis tau statistic (Siegal, 1956) was also used.

To quantitatively investigate the relation between moisture content and lithology, simple linear regressions were calculated, using the least squares method, with moisture content as the dependent variable and percent sand, percent silt, percent clay, percent shrinkage and specific surface as alternate independent variables. The analysis was conducted using the eighteen representative samples selected for laboratory analysis.

Sampling plots were initially set up so that the effects of aspect on moisture content may be investigated. East facing and west facing plots were selected with duplicate lithologies and where possible similar slope angles. To quantitatively investigate the effects of aspect the Mann-Whitney U test (Hammond and McCullagh, 1974) was used. The Student's t-test was not used because of sensitivity to non-normality of data (Snedecor and Cochran, 1967). The non-parametric Mann-Whitney U method tests the null hypothesis that the two samples being investigated (in this case one sample with east facing aspect and one sample with west facing aspect) are drawn from the same population. The chosen level of significance was 0.05, i.e. if the Mann-Whitney U is significant at a level higher (less than 0.05) than the 0.05 level, then, it can be concluded that the two samples investigated are from different populations.

In order to investigate factors related to the percent shrinkage of the various materials regression analyses were carried out with percent shrinkage as the dependent variable and percent sand, percent silt, percent clay and specific surface as alternate independent variables.

CHAPTER V

RESULTS

5.1 INTRODUCTION

In this chapter the results of the laboratory tests on samples are presented and discussed. This is followed by a discussion of the various analyses performed on these data.

5.2 RILL COUNTS

A total of 177 rills were counted on the sandstone surfaces and 54 were counted for the adjacent shale surface with respective mean values of 17.7 (standard deviation of 3.3) and 5.4 (standard deviation of 2.6), Table 5.1. Student's t-test showed these means were significantly different at the 0.001 level. It was therefore concluded that rills are more frequent on the sandstone lithology. This suggests that surface runoff is more dominant on the sandstone lithology than on the shale

Table 5.1 RILL COUNTS OVER TWO METRE SLOPE WIDTHS

Clay Cemented Sandstone	Shale
21	1
15	4
16	6
16	5
21	6
11	3
21	4
17	9
20	9
19	7

lithology. The possibility exists that these figures may merely indicate concentrated flow on the sandstone and unconcentrated flow on the shale. In view of the micro-topography of the shale due to desiccation cracking, however, this is unlikely. Since these two lithologies are juxtaposed they are obviously receiving the same inputs of rainfall. This leads to the conclusion that the shale has a greater infiltration rate allowing less surface runoff to occur than the sandstone. This compares with the findings of Schumm (1956a, 1956b) and Schumm and Lusby (1964), in investigating the Brule (cf. to the sandstone) and Chadron (cf. to the shale) formations. Although it appears that more water is infiltrated into the shale, it does not necessarily mean that it is held in the "soil." It may run off as subsurface flow either as throughflow, or more commonly probably as pipeflow. In terms of erosion this would imply greater subsurface erosion on the shale lithology than on the sandstone. Erosion studies carried out by Campbell (1970, 1974, 1978) since 1968 found that erosion rates between plots for the shale lithologies showed greater standard deviations than the sandstone lithologies. Greater standard deviations would be expected for subsurface erosion than for the surface removal of material from a smooth surface and this tends to confirm the conclusion of greater subsurface water flow on the shale relative to the sandstone.

5.3 MOISTURE CONTENT

The mean moisture content for each sampling site along with their standard deviations, standard errors, percent standard errors and 95 percent confidence limits are shown in Table 5.2. Mean moisture content

Table 5.2 MOISTURE CONTENTS OF VARIOUS LITHOLOGIES

Sample	Number in Sample	Mean θdw	Standard Deviation	Percent Standard Error	95 % Confidence Limits
1A	13	0.28	0.22	21.79	0.158 - 0.402
1B1	13	0.27	0.18	18.52	0.170 - 0.370
1B2	12	0.28	0.20	19.64	0.170 - 0.390
1C1	13	0.46	0.38	22.83	0.250 - 0.670
1C2	13	0.39	0.35	24.87	0.190 - 0.590
1D	13	0.52	0.39	20.77	0.300 - 0.740
2AN	7	0.31	0.20	24.19	0.160 - 0.460
2BN	7	0.22	0.11	18.64	0.140 - 0.300
2CN	7	0.30	0.14	17.67	0.190 - 0.410
2DN	7	0.25	0.16	24.00	0.130 - 0.370
2AO	4	0.19	0.21	55.26	0.020 - 0.400
2BO	5	0.39	0.29	33.08	0.130 - 0.650
2CO	5	0.50	0.37	3.40	0.470 - 0.530
4A	8	0.16	0.10	21.88	0.090 - 0.230
4B	8	0.16	0.16	35.63	0.050 - 0.270
5A	8	0.17	0.10	20.59	0.100 - 0.240
5B	8	0.13	0.14	37.69	0.030 - 0.230
6A	6	0.20	0.14	28.50	0.090 - 0.310
6B	6	0.19	0.17	36.32	0.050 - 0.330
6C	6	0.17	0.17	40.59	0.030 - 0.310
7A	8	0.14	0.10	25.00	0.070 - 0.210
7B	7	0.13	0.08	23.08	0.070 - 0.190
7C	8	0.09	0.08	31.11	0.030 - 0.150
8A	7	0.86	0.11	4.88	0.780 - 0.940
8B	7	0.19	0.16	31.58	0.070 - 0.310
8C	7	0.22	0.14	24.09	0.110 - 0.330
9A	8	0.11	0.09	29.09	0.050 - 0.170
9B	8	0.10	0.11	39.00	0.020 - 0.180
9C	8	0.33	0.31	33.33	0.310 - 0.350
10A	6	0.70	0.05	28.57	0.660 - 0.740
10B	6	0.92	0.05	21.74	0.880 - 0.960
10C	6	0.12	0.07	24.17	0.060 - 0.180
10D	6	0.15	0.08	22.00	0.080 - 0.220
11A	6	0.06	0.05	33.33	0.020 - 0.100
11B	6	0.10	0.07	29.00	0.040 - 0.160
11C	6	0.12	0.13	44.17	0.010 - 0.220
11D	6	0.08	0.08	41.25	0.010 - 0.150

ranges from 0.06 to 0.92 with standard deviations of 0.05 for both figures. The maximum percent standard error was 55.26 percent for yellow bentonite on the crest slope and the minimum was 3.40 percent also for yellow bentonite but on the midslope section of the same plot.

From inspection of the mean moisture contents and their associated 95 percent intervals three groups of precision appear to be present (Fig. 5.1). Loosely, these three groups may be defined as those sites having very large confidence intervals (covering a range greater than 0.25); those with medium confidence intervals (0.10 to 0.20); and those with a high degree of precision or small confidence intervals (less than 0.10). Moisture relationships will be discussed in detail in a later section of this chapter.

5.4 RAINFALL

Rainfall amount was calculated as the arithmetic mean of the four gauges located in and adjacent to the study area (Fig. 4.1). The mean rainfall for the study period was 7.5 mm with a standard deviation of 4.5 mm. A total of 13 rainfall events were investigated (Table 5.3). Percent standard errors between the gauges ranges from 0 to 48.9 percent indicating a very high degree of variability within the basin. This extreme variability of rainfall seems typical of semiarid environments (Osborn and Renard, 1970; Sharon, 1970), and it poses certain problems to field-based investigations of the type reported here (see 6.2).

FIG. 5.1 CONFIDENCE INTERVALS OF MEAN MOISTURE CONTENTS FOR VARIOUS LITHOLOGIES

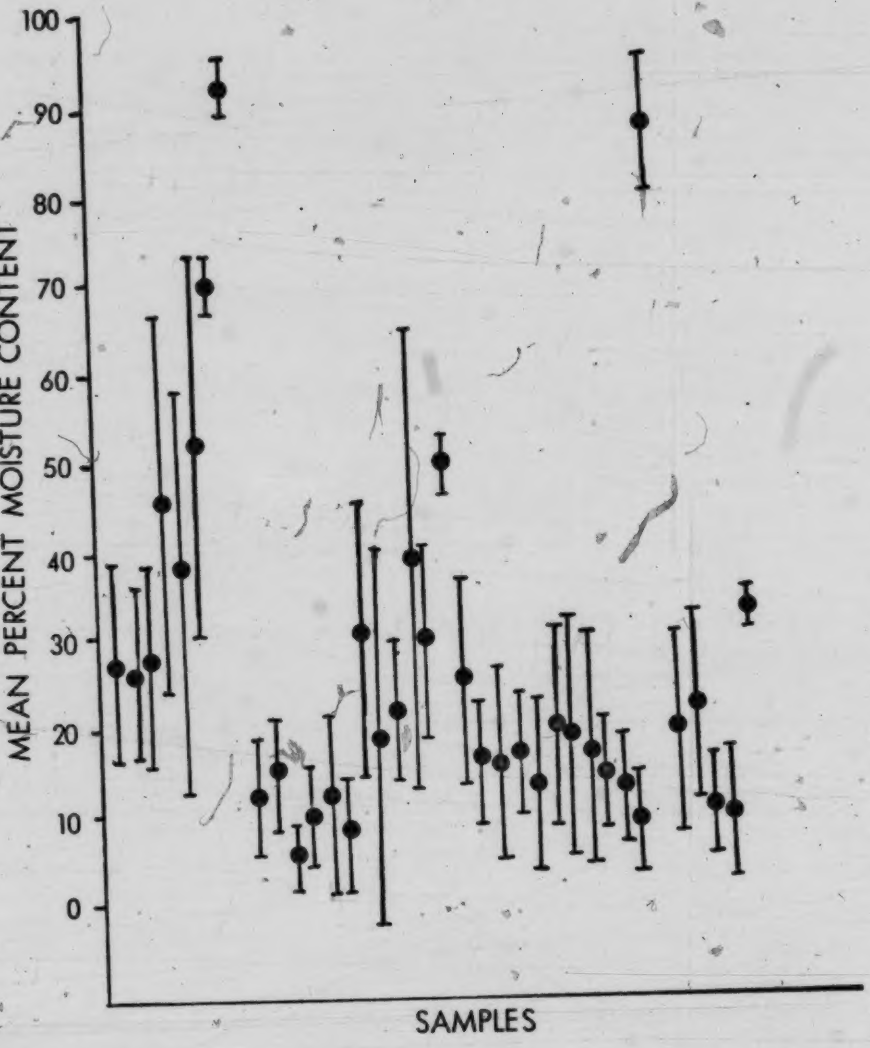


Table 5.3 RAINFALL DATA FOR THE STUDY PERIOD

Storm Number	Rainfall Amount (mm)	Mean Intensity (mm/hr)	Maximum Intensity (mm/hr)	Duration of Storm (hrs)
1	6.57*			
2	10.00*			
3	10.00*			
4	19.37	1.23	2.12	12.05
5	10.25	0.52	0.64	25.00
6	2.63	0.94	1.66	5.19
7	5.16	1.98	3.18	2.23
8	4.87	1.66	2.12	2.74
9	6.90	3.01	6.00	5.25
10	3.96	4.04	5.08	1.00
11	4.00	6.84	15.24	4.17
12	2.62	1.14	1.50	3.00
13	10.81	7.78	30.00	12.00

*Intensity and duration data not available.

5.5 RAINFALL INTENSITY AND DURATION

Rainfall intensity and duration records were available from the fourth rainfall event to the completion of the field season (16th May to 12th August, 1977). The average intensity for each event for which moisture samples were taken was used in data analysis and was calculated as the arithmetic mean of the mean intensity from each chart from the four gauges in the study area. Mean storm duration was calculated similarly.

The mean intensity for the storms investigated was 3 mm/hr with a standard deviation of 2.6 mm/hr. Mean intensity for each storm ranged from 0.5 mm/hr to 8 mm/hr (Table 5.3) which does not seem high (Gregory and Walling, 1973; Toogood, 1977). Maximum intensity is also shown in

Table 5.3. This ranged from 1 mm/hr over four hours up to 30 mm/hr received in about 10 minutes.

The mean storm duration for the events investigated was 7.20 hours with a standard deviation of 7.42 hours. The mean for each storm ranged from one hour to 25 hours (Table 5.3).

An intensity-duration curve from data covering the entire record period (16th May to 12th August, 1977) gave an inverse shape to that normally expected in a drainage basin (Gray, 1970; Gregory and Walling, 1973), Fig. 5.2. This may be due to insufficient data, i.e. too short a record.

5.6 PARTICLE SIZE ANALYSIS

The 18 representative samples showed a wide range of variability of sand, silt and clay fractions (Fig. 5.3). Percent sand ranged from three percent to 55 percent with a mean of 19.67 and standard deviation of 15.95; percent silt ranged from 23 percent to 80 percent with a mean of 46.48 and a standard deviation of 13.00; while percent clay ranged from eight percent to a 63 percent with a mean of 34.21 and a standard deviation of 17.11.

These were plotted on the soil texture triangle and most values, as expected, fell outside the "normal" range of agricultural soils, falling in the silty clay loam to heavy clay region (Fig. 5.4). The soil investigated plots as a loam while three samples from the highly rilled sandy surfaces plot as a clay loam (8A), sandy clay loam (1Ao), and a silt loam (1A).

FIG. 5.2 INTENSITY-DURATION CURVE

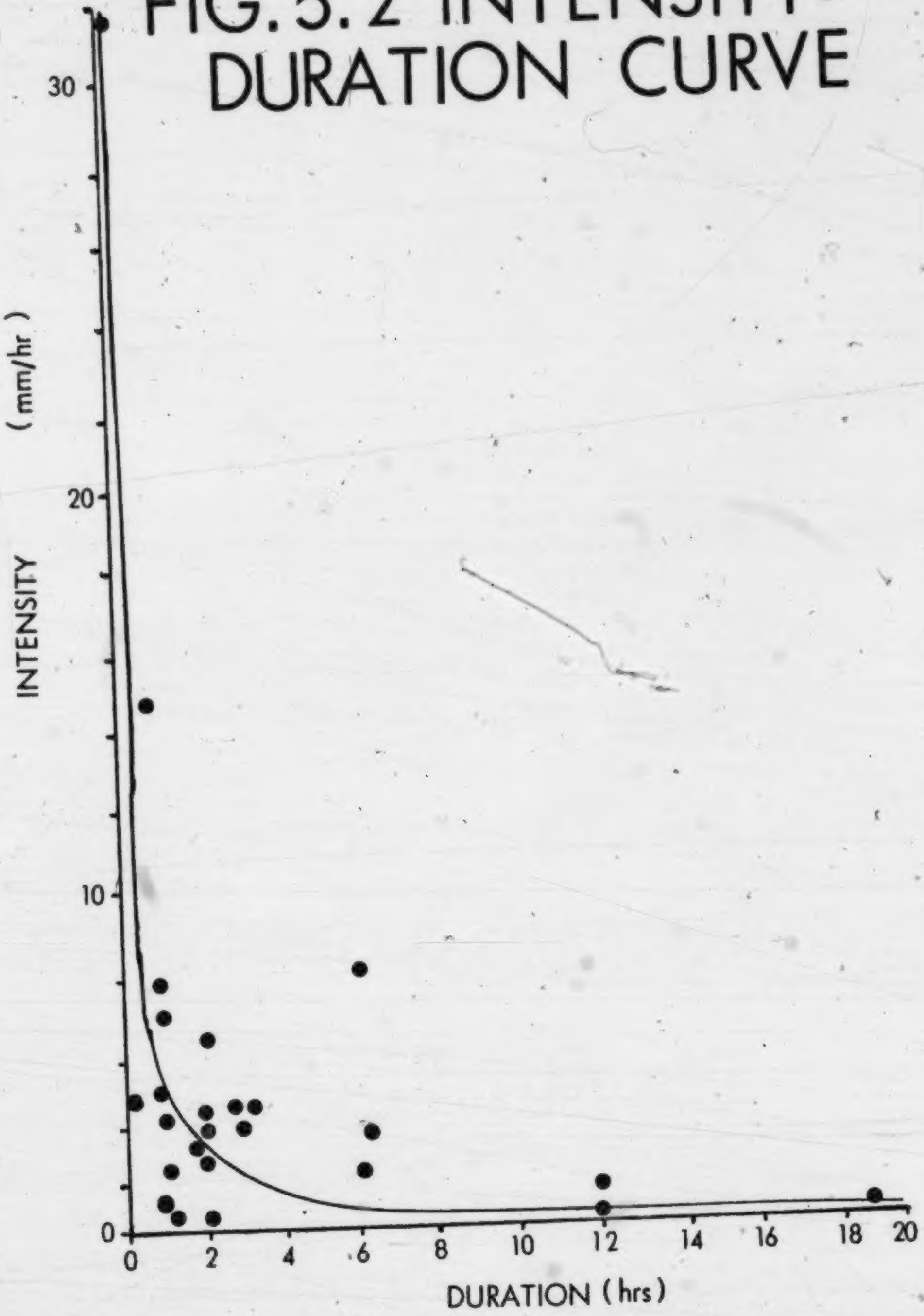
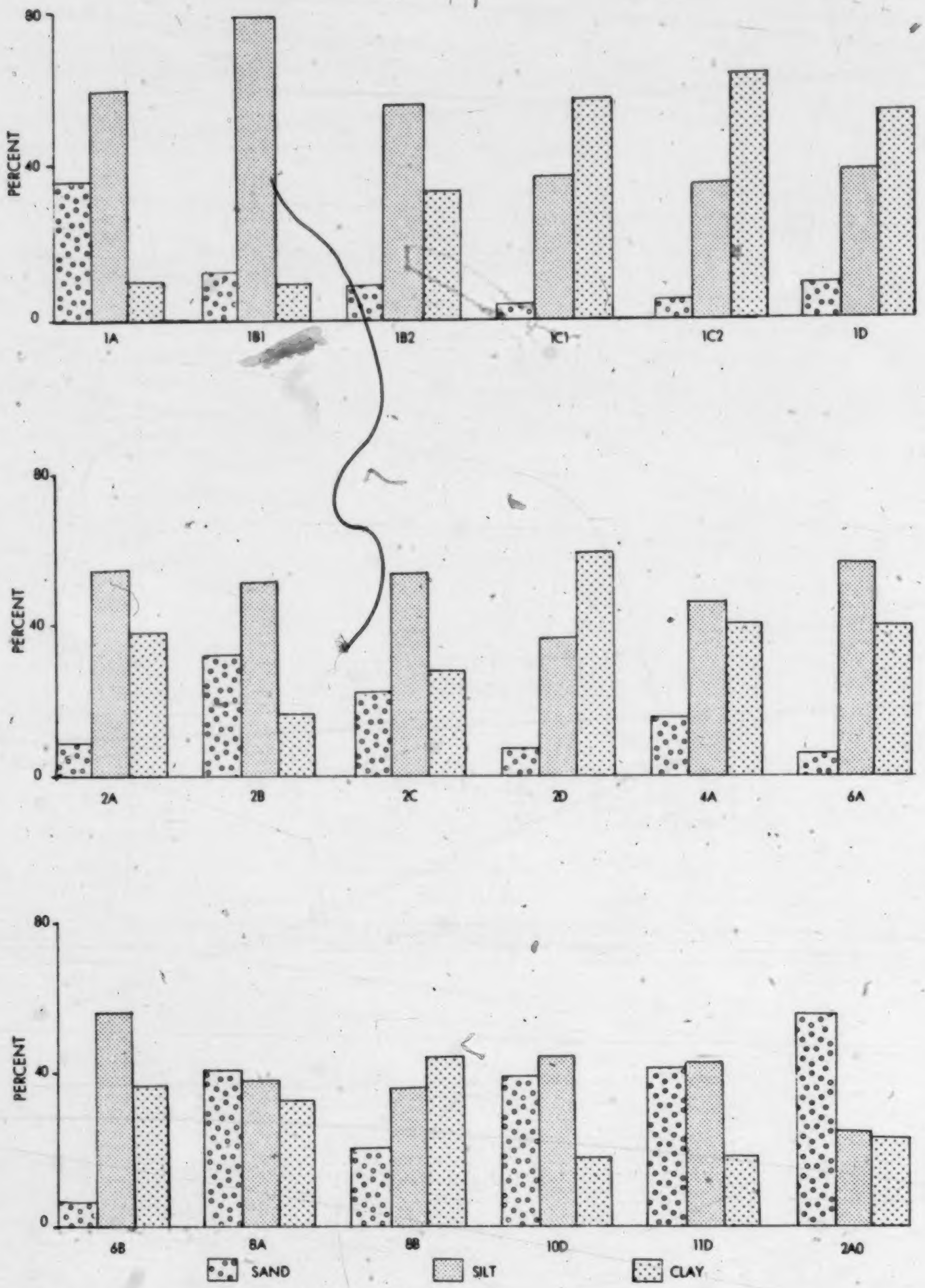


FIG. 5.3 HISTOGRAMS of PARTICLE SIZE FRACTIONS



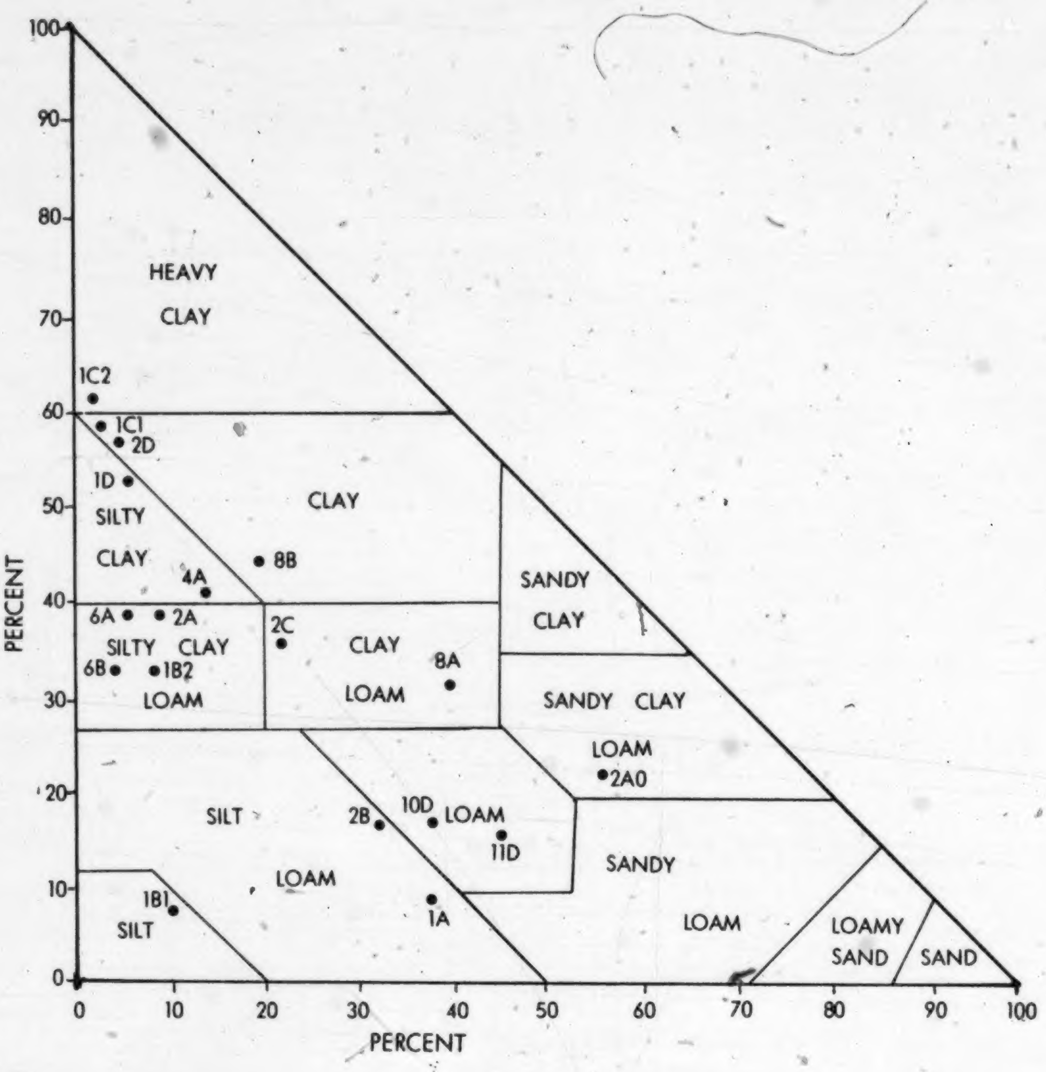


FIG. 5. 4 SOIL TEXTURE TRIANGLE

5.7 SPECIFIC SURFACE

The duplicate specific surface measurements showed close agreement so the mean of the two measurements for each sample was taken as the actual specific surface value. These ranged from 35 m^2/g for the soil material up to 328 m^2/g for grey clay (Table 5.4). The mean was 200.87 m^2/g with a standard deviation of 79.76. The higher specific surface reflects the quantity of montmorillonite clay in the sample (Baver, _et al._, 1972). Least squares regression analysis between specific surface and total percent montmorillonite in the sample gave an r^2 value of 0.59, significant at the 0.01 level. Since specific surface gives an indication of the amount of water each sample is able to contain, high

Table 5.4 SPECIFIC SURFACE OF REPRESENTATIVE SAMPLES

Sample Number	Specific Surface (m^2/g)
1A	186.16
1B1	218.53
1B2	237.62
1C1	313.95
1C2	274.60
1D	327.92
2A	182.31
2B	160.26
2C	216.28
2D	200.05
4A	128.95
6B	174.65
6A	133.45
8A	200.66
8B	203.43
10D	46.11
11D	34.49
2A0	130.15

values indicate a high degree of moisture capacity and low values indicate a lower degree of moisture capacity.

5.8 CLAY MINERALOGY

The dominant clay mineral present in all but the soils is montmorillonite. Percent montmorillonite in the clay fraction ranged from eight percent for a soil to 100 percent for reddish clay. Mean percent montmorillonite was 23 percent with a standard deviation of 15.4 percent. Illite was the next most prevalent clay mineral being the dominant mineral in both the true soils tested. Percent illite ranged from zero for reddish clay to 66 percent for soil. Mean percent illite was eight percent with a standard deviation of 5.3 percent. Smaller amounts of kaolinite were found, ranging from zero for rilled sandstone nd grey and reddish clays to 20 percent for a soil. Mean percent kaolinite was three percent with a standard deviation of 3.6 percent (Table 5.5).

Although the clay minerals present in the clay fraction expressed as a proportion of the clay fraction are useful, for practical purposes, it is more appropriate to consider the actual amount of each clay mineral expressed as a proportion of the entire sample and not just the clay fraction. These values were calculated (Table 5.6) and it was found that percent montmorillonite in the total sample ranged from one percent for a soil up to 58 percent for reddish grey clay, with a mean of 23 percent and standard deviation of 15.37. Percent illite as percentage of the sample ranged from zero percent for a reddish clay to 16 percent for grey clay with a mean of eight percent and standard deviation of 5.33. Percent kaolinite as a percentage of the total

Table 5.5 CLAY MINERALOGY AS PERCENT OF CLAY FRACTION

Sample Number	Percent Montmorillonite	Percent Illite	Percent Kaolinite
1A	93	7	0
1B1	100	0	0
1B2	94	6	0
1C1	83	17	0
1C2	91	9	0
1D	72	2	26
2A	42	35	23
2B	65	28	8
2C	69	20	11
2D	67	24	9
4A	50	40	7
6A	64	31	5
6B	52	40	8
8A	78	17	5
8B	76	22	3
10D	8	66	15
11D	11	63	20

Table 5.6 CLAY MINERALOGY AS PERCENT OF TOTAL SAMPLE

Sample Number	Percent Montmorillonite	Percent Illite	Percent Kaolinite
1A	9	1	0
1B1	8	0	0
1B2	32	2	0
1C1	45	10	0
1C2	58	6	0
1D	39	1	14
2A	16	13	9
2B	10	4	1
2C	18	5	3
2D	39	14	6
4A	20	16	3
6A	25	12	2
6B	19	15	3
8A	25	6	2
8B	33	10	1
10D	1	11	3
11D	2	11	4

sample ranged from zero percent for rilled sandstone, reddish grey and reddish clay to 14 percent for grey clay with a mean of three percent and a standard deviation of 3.60.

On inspecting a histogram (Fig. 5.5) of actual percent clay minerals present in the whole sample, six groups of mineralogies can be distinguished. Those with moderate amounts of montmorillonite but small amounts of illite and kaolinite (Type A); those with large amounts of montmorillonite and small amounts of illite and kaolinite (Type B); a sample with high amount of montmorillonite, small amount of illite but moderate amount of kaolinite (Type C); those with moderate amounts of all three minerals but decreasing in amount from montmorillonite to illite to kaolinite (Type D); those with relatively high amounts of montmorillonite, moderate amounts of illite and moderate to small amounts of kaolinite (Type E); and those with relatively high amounts of illite but small amounts of montmorillonite and kaolinite (Type F). Type F mineralogy is from the two soils and is a typical clay mineralogy for soils of arid and semiarid areas (Grim, 1968).

5.9 PERCENT SHRINKAGE

The percent shrinkage measured in the 18 representative soils range from 7 percent for a grey clay to 51 percent for grey bentonite with a mean of 32.8 percent and standard deviation of 13.98 (Table 5.7). As may be expected, the sandy soils show a lower degree of shrinkage than most of the clay-rich materials. One sample (1B1) shows 45 percent shrinkage but contains only eight percent clay (Fig. 5.3). All of the clay is montmorillonite, however, which probably explains this amount

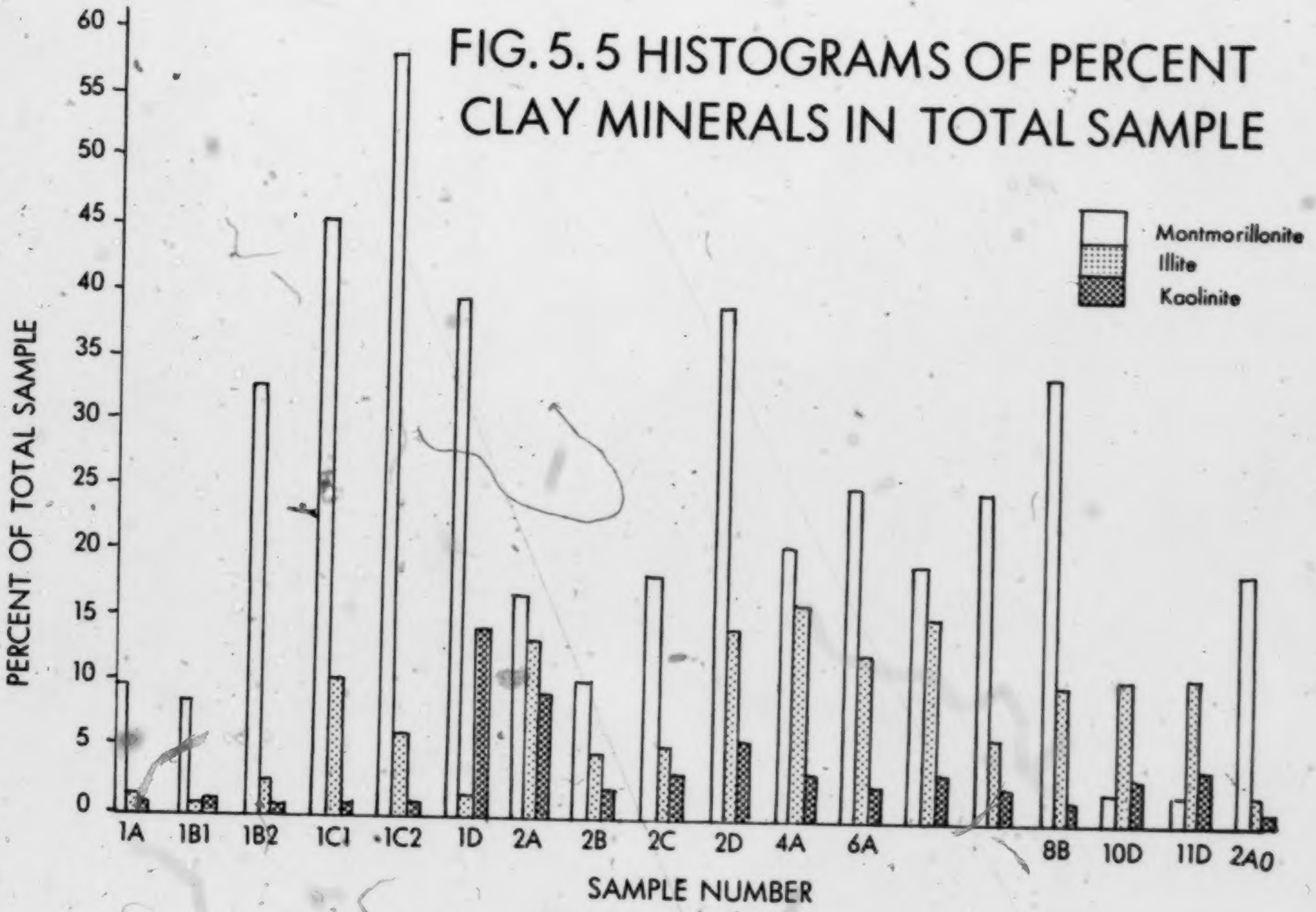


FIG. 5.5 HISTOGRAMS OF PERCENT CLAY MINERALS IN TOTAL SAMPLE

Table 5.7 PERCENT SHRINKAGE AND APPROXIMATE ATTERBERG LIQUID LIMITS

Sample Number	Percent Shrinkage	Liquid Limit
1A	28.6	43
1B1	45.4	68
1B2	17.0	59
1C1	32.9	80
1C2	47.2	67
1D	50.5	61
2A	29.6	56
2B	36.2	45
2C	42.3	57
2D	47.8	58
4A	30.7	42
6B	35.7	46
8A	39.6	68
8B	45.7	71
10D	7.5	49
11D	10.3	51
2A0	37.5	75
2C0	7.4	83

of shrinkage.

In performing the measurements of shrinkage the samples were wetted to their approximate liquid limits before being dried and allowing the shrinkage to occur. The moisture contents of these samples on a dry weight basis may therefore be regarded as approximate indications of the Atterberg liquid limits of the samples. These values range from 43 percent to 83 percent with a mean of 60 percent and standard deviation of 13 percent (Table 5.7). If anything these data probably overestimate the true liquid limits of the samples and should therefore be treated with caution.

5.10 NORMALITY OF THE DATA

Tests carried out to examine the normality of data showed that all moisture content data pooled gave a high positive kurtosis and also a positive skewness indicating that the data do not follow a normal distribution (Table 5.8). Alternatively for each sampling site individually, four sites showed a positive skew while eight sites showed a negative kurtosis value (Table 5.8). For all other sites, only slight skewness and negative kurtosis values (three sites showed slight positive kurtosis values) were evident. Although this raises the question regarding the normality of the data, it was concluded that the hypothesis that the moisture contents for individual sampling sites follow a normal distribution should not be rejected since the majority of the sites showed only slight deviation of skewness and kurtosis from the normal distribution.

Similar conclusions were reached from the skewness and kurtosis values for the lithological variables (Table 5.9). Percent clay showed a definite negative kurtosis value indicating non-normality of the data. Percent sand, percent silt, percent shrinkage and specific surface, however, all showed only slight negative kurtosis. Percent shrinkage and specific surface show a slight negative skew while percent sand, percent silt, and percent clay show a slight positive skew. It was concluded that, although the slight skewness and kurtosis values could lead one to be suspicious of the normality of the data, the hypothesis that percent shrinkage, specific surface and percent sand, silt and clay follow a normal distribution should not be rejected again, since only slight aberrations from normality were found.

Table 5.8 SKEWNESS AND KURTOSIS OF MOISTURE DATA

Sample Number	Skewness	Kurtosis
1A	0.946	0.498
1B1	0.210	-1.034
1B2	0.473	-0.531
1C1	0.553	-0.478
1C2	0.344	-1.316
1D	0.182	-1.239
2AN	-0.052	-0.956
2BN	0.706	-0.545
2CN	0.082	-1.345
2DN	0.217	-1.294
2AO	0.120	-1.816
2BO	0.003	-1.662
2CO	0.027	-1.696
4A	-0.017	-1.278
4B	1.179	-0.266
5A	0.328	-1.201
5B	0.848	-0.627
6A	-0.085	-0.874
6B	0.095	-1.370
6C	0.541	-0.860
7A	-0.256	-1.128
7B	-0.355	-0.995
7C	0.013	-1.564
8A	1.114	-0.098
8B	0.585	-0.637
8C	1.227	0.361
9A	1.032	0.018
9B	0.567	-0.743
9C	1.489	0.930
10A	-0.369	-1.023
10B	-0.629	-0.184
10C	-0.067	0.264
10D	-0.627	-1.445
11A	-0.238	-1.680
11B	0.280	-0.665
11C	0.979	-0.229
11D	0.307	-0.990
Pooled Data	1.719	3.239

Table 5.9 SKEWNESS AND KURTOSIS FOR LITHOLOGIC VARIABLES

Parameter	Skewness	Kurtosis
% Shrinkage	-0.697	-0.700
% Sand	0.772	-0.684
% Silt	0.576	-0.476
% Clay	0.123	-1.067
Specific Surface	-0.248	-0.152

Unlike the usual negatively skewed distribution of rainfall amount, rainfall data for the badlands which were investigated, show a positive skew (Table 5.10). The data also exhibited a positive kurtosis value leading to the expected conclusion that rainfall does not follow a normal distribution. Reasons for a positive skew are probably twofold. Firstly, not all storms recorded for the period 16th May to 12th August when records were available were included; only those storms investigated (i.e. those for which moisture samples were taken) were included in the analysis. Secondly, the data are from only a brief period of the rainfall season and the season may have been an atypical one.[1]

Table 5.10 SKEWNESS AND KURTOSIS FOR RAINFALL

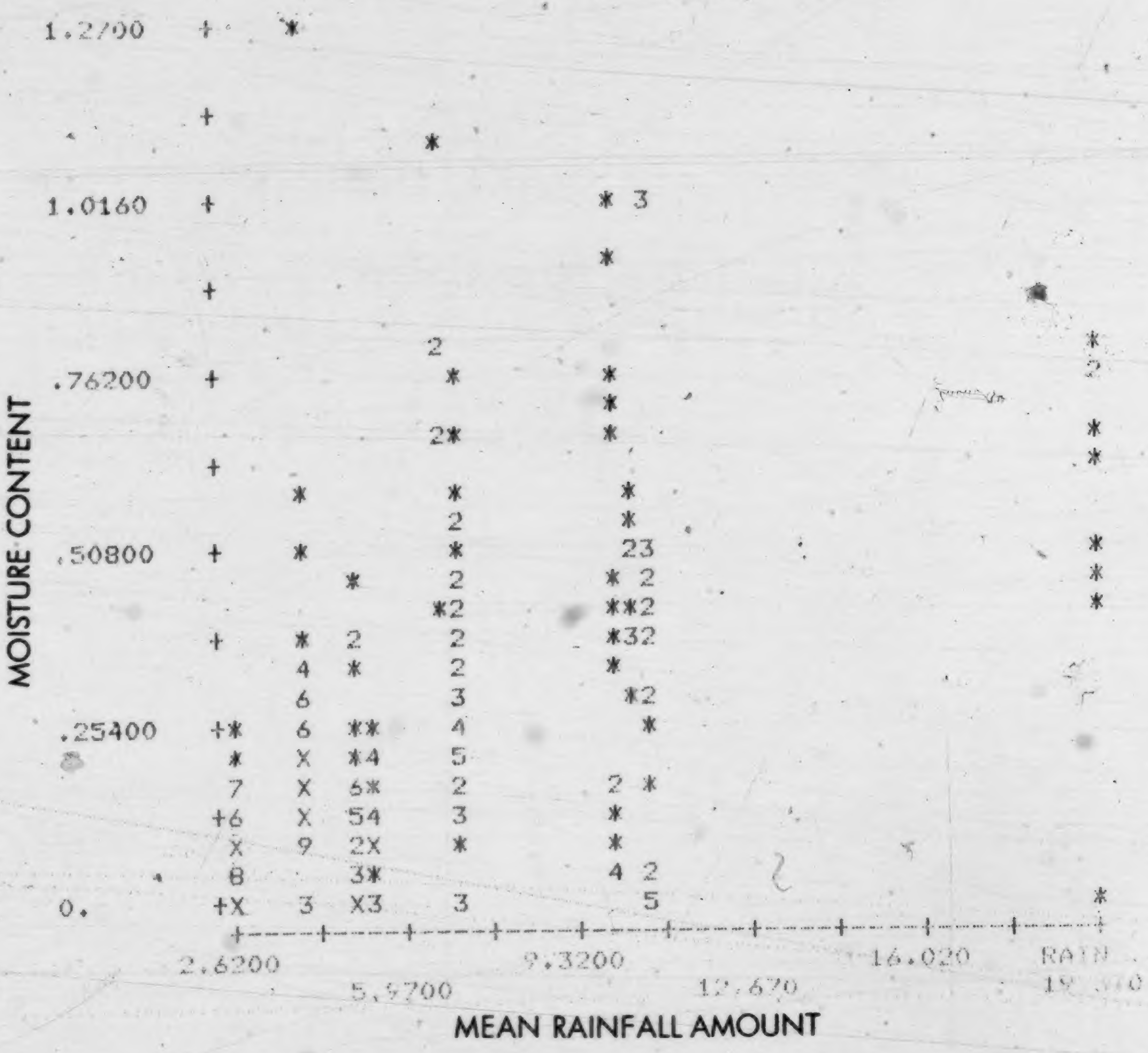
	Skewness	Kurtosis
Rainfall Amount	1.244	1.249

5.11 RELATION BETWEEN MOISTURE CONTENT AND RAINFALL

Moisture samples collected and treated in the laboratory as outlined in Chapter IV show a general tendency to increase with the rainfall amount up to about 12 mm and the slope of the curve appears to decrease and flatten (Fig. 5.6). Interpretation of the graph, however, is hindered by the small amount of data for rainfalls greater than 12 mm (only data for one storm was available in this range). To statistically test the hypothesis that there is a straight line relationship between moisture content and rainfall amount a simple linear regression was carried out with moisture content as the dependent and rainfall amount less than 12 mm as the independent variables. This yielded an r^2 value of 0.26, significant at the 0.01 level, indicating that only 26 percent of the variation of moisture content could be explained by the variation in rainfall amount. The regression equation was

$$\theta dw = 0.0312 + 0.324 \text{ (rainfall amount less than 12 mm)}$$

Since such a wide scatter of data is evident it was concluded that investigation of moisture contents relating to each sample plot may be more fruitful. From a total of 38 sample sites, however, only 17 yielded significant regression equation (Table 5.11), with the coefficient of determination (r^2) ranging from 0.25 to 0.88 and the significance levels ranging from 0.08 to 0.001. The equations show the intercept values ranging from -0.220 to 0.157 and the slopes of the lines from -0.021 to 0.108. These data should be interpreted with caution since sample sizes are in some cases quite small (n = 5 is the smallest). The wide degree of scatter probably reflected the small sample sizes and therefore sites were grouped on the basis of lithology to pool data of like origin and hopefully improve regression estimates. The hypothesis



5.6 Moisture content-rainfall amount relation for pooled data.

Table 5.11 SIMPLE REGRESSION EQUATIONS FOR INDIVIDUAL SITES OF MOISTURE CONTENT VERSUS RAINFALL LESS THAN 12 mm

Sample	R^2	Significance
1A	0.20	0.1460
1B1	0.15	0.2122
1B2	0.08	0.3863
1C1	0.02	0.6299
1C2	0.18	0.1629
1D	0.46	0.0148
2AN	0.43	0.1098
2BN	0.73	0.0141
2CN	0.57	0.0503
2DN	0.09	0.5061
2AO	0.51	0.4952
2BO	0.38	0.3837
2CO	0.38	0.3837
4A	0.58	0.0273
4B	0.79	0.0031
5A	0.52	0.0421
5B	0.32	0.1451
6A	0.69	0.0395
6B	0.70	0.0382
6C	0.004	0.9020
7A	0.41	0.0864
7B	0.20	0.3165
7C	0.17	0.3047
8A	0.02	0.7387
8B	0.53	0.0640
8C	0.58	0.0456
9A	0.07	0.5321
9B	0.001	0.9217
9C	0.87	0.0006
10A	0.35	0.2198
10B	0.62	0.0635
10C	0.21	0.3624
10D	0.08	0.5772
11A	0.24	0.3270
11B	0.60	0.0724
11C	0.59	0.0756
11D	0.43	0.1601

being tested thus became that moisture contents of similar "soil" types showed a straight line relationship with rainfall amounts less than 12 mm. This was to enable a comparison between soil types on the basis of moisture content-rainfall amount relationships. It was expected that each soil type would have a unique relationship with rainfall amount. From 11 groups of soil types, only three regressions were significant at the 0.01 level (Table 5.12). The coefficients of determination, however, were low with a maximum of 0.29 indicating only 29 percent of the variation of moisture content was due to the variation in rainfall amount. This value was for the soil group in the lower portion of the basin. It was obvious therefore, that little could be explained by rainfall amount alone.

Table 5.12 SIMPLE REGRESSIONS FOR SAMPLING SITES GROUPED FOR LITHOLOGY

Groups	R^2	Significance
10A thru 11A	0.29	0.0001
2D, 1C1, 1C2, 1D	0.14	0.0119
2BN, 1A	0.27	0.0224
6B, 7C, 6C, 7D	0.02	0.4728
8B, 8C, 9B, 9C	0.34	0.0008
2BO, 2CO	0.44	0.0367
1B2, 2AN, 6A, 7A, 7B	0.29	0.0003
4A, 4B, 5A, 5B	0.52	0.0000
8A, 9A	0.01	0.7777
2AO	0.51	0.4952
2CN	0.57	0.0503

5.12 EFFECTS OF INTENSITY

In order to test the hypothesis that soil moisture increases linearly with intensity of rainfall less than 12 mm further regression analysis was conducted. The hypothesis that moisture content increases linearly with duration of rainfall less than 12 mm was tested similarly. Total regressions with all data pooled showed significant relationships for both intensity and duration. The significance, however, is probably due to the large amount of data since the coefficients of determination were only 0.06 and 0.18 for moisture content versus intensity and moisture content versus duration respectively.

Multiple regressions were calculated in an attempt to improve estimates showing moisture content as a function of rainfall amount, intensity and duration. For the total regression with all data pooled for rainfall less than 12 mm, least squares regression analysis yielded an r^2 of 0.32, significant at the 0.01 level. The multiple regression, as would be expected, shows some improvement over separate regressions with each variable (Table 5.13). It is interesting to note that rainfall duration has a higher level of explanation than does rainfall intensity, suggesting, contrary to Horton's (1945) concepts, that duration of rain is more important than intensity in the infiltration-runoff process on these materials. This is discussed in more detail in Chapter VI.

Since little explanation could be gained from the multiple regression using all data, regressions were carried out on each sampling site testing the hypothesis that moisture content is a linear function of rainfall amount less than 12 mm, intensity and duration for each sampling site. Only four equations out of 38 calculated were signifi-

Table 5.13 MULTIPLE REGRESSIONS OF MOISTURE CONTENT VERSUS RAINFALL LESS THAN 12 mm, INTENSITY AND DURATION

Sample	R^2	Significance
Pooled	0.26	0.0000
1A	0.89	0.0079
1B1	0.45	0.3600
1B2	0.53	0.2485
1C1	0.42	0.3975
1C2	0.71	0.0806
1D	0.77	0.0484
2AN	0.43	0.5867
2BN	0.75	0.1918
2CN	0.58	0.3970
2DN	0.44	0.5810
2AO	–	– *
2BO	–	– *
2CO	–	– *
4A	0.73	0.1265
4B	0.88	0.0244
5A	0.62	0.2295
5B	0.72	0.1300
6A	0.92	0.1227
6B	0.79	0.2925
6C	0.28	0.8536
7A	0.55	0.3206
7B	0.42	0.5982
7C	0.40	0.5144
8A	0.73	0.2151
8B	0.63	0.3351
8C	0.68	0.2781
9A	0.40	0.5148
9B	0.21	0.7931
9C	0.92	0.0119
10A	0.50	0.6419
10B	0.71	0.3989
10C	0.49	0.6596
10D	0.77	0.3220
11A	0.66	0.4613
11B	0.71	0.4063
11C	0.74	0.3646
11D	0.63	0.4947

* Rainfall intensity and duration data not available

cant at the 0.05 level (Table 5.14).

Table 5.14 SEPARATE REGRESSIONS OF MOISTURE CONTENT ON RAINFALL INTENSITY AND DURATION FOR POOLED DATA

Variable	R^2	Significance
Intensity	0.06	0.0001
Duration	0.18	0.0000

Again data were grouped lithologically to increase sample sizes and of 11 groups, seven equations were significant (Table 5.15). The

Table 5.15 MULTIPLE REGRESSIONS OF MOISTURE CONTENT VERSUS RAINFALL AMOUNT, INTENSITY AND DURATION GROUPED LITHOLOGICALLY

Groups	R^2	Significance
10A thru 11D	0.39	0.0001
2D, 1C1, 1C2, 1D	0.30	0.0073
2B, 1A	0.76	0.0073
6B, 6C, 7D, 7C	0.09	0.5720
8B, 8C, 9B, 9C	0.36	0.0083
1B2, 2AN, 6A, 7A, 7B	0.50	0.0000
4A, 4B, 5A, 5B	0.64	0.0000
8A, 9A	0.37	0.1552
2AO	-	- *
2CO	-	- *
9C	0.92	0.0119

* Rainfall intensity and duration data not available

r^2 values ranged from 0.30 to 0.92 with associated significance levels of 0.01 for grey clay and yellow bentonite respectively, suggesting a considerable improvement over the previous analysis on grouped data with rainfall alone.

5.13 RELATION BETWEEN MOISTURE CONTENT AND LITHOLOGY

None of the correlations between moisture content and lithology were significant implying that no relation existed between type of lithology and moisture content. This certainly seems surprising in view of what is normally understood about soil behaviour (Hillel, 1971; Baver, et al., 1972) and also after consideration of research more closely related to the present study (e.g. Bryan, et al., 1978; Grim, 1968). No relation could be found between decreasing significance of r^2 values from any of the analyses and any of the lithological variables investigated. This suggests that for these materials lithological properties have little association with moisture-rainfall relationships.

5.14 ASPECT

Sampling sites were relegated to two samples of the population, one sample with easterly aspect and one with westerly aspect. Calculation of the Mann-Whitney U statistic yielded a value of 8330 significant at the 0.006 level. The null hypothesis was therefore rejected and it was concluded that the effect of aspect was real and two populations were present in the data. The median test led to similar conclusions. Since it is apparent that two populations are present in the data, regression analyses relating moisture content to rainfall amount were carried out on the separate populations, grouping according to lithology. Multiple regressions were also calculated for moisture content and rainfall amount, intensity and duration. No improvement, however, was evident in significance levels or coefficients of determination over the previous analyses.

The probable reason for the differences in aspect between the plots is that the prevailing winds are from the southwest. No records, however, were taken of the wind direction during storm events. One aspect does not appear to be consistently higher in moisture content than the other. In view of the highly localised nature of the rain catch measured on the few occasions where an intensive gauge network was available (Bryan, *et al.*, 1978) this is perhaps not surprising. Obviously more intensive investigations of micro-climate are required before this behaviour may be fully explained.

An attempt to use infiltration rings in the field failed completely because of the problem of lateral flow. The rings were left in place, however, and it was noted that even after three months, the material inside the rings which had been saturated, had not developed a "popcorn" layer comparable to the surrounding material of the same nature which had been wetted in the same manner at the same time. This further emphasises the importance of micro-climatic factors. The rings offered some shade from the sun, but more important sheltered the small area from the wind thus altering the rate of desiccation of the material.

5.15 SHRINKAGE-LITHOLOGY RELATIONSHIPS

Only one of the regressions (percent shrinkage versus percent silt) was significant at the 0.05 level. Since the r^2 value was only 0.04 this equation was not considered to be of any value. Percent shrinkage was not related to specific surface either. A multiple regression analysis was attempted with all independent variables together but this analysis was not significant.

After further consideration a new variable, percent fines, was established, this being the sum of percent clay and percent silt for each sample. A linear regression with percent shrinkage as the dependent and percent fines as the independent variables was significant at the 0.05 level but the r^2 value was only 0.23 indicating only 23 percent of the variation in shrinkage was due to the variation in percent fines. One would expect a much higher degree of explanation since it is the clay which is responsible for the shrink - swell phenomena in soils (Grim, 1968; Baver, _et al._, 1972). The wide degree of scatter may perhaps partially be explained by the range of clay minerals present in various amounts in each clay fraction.

The degree of shrinkage was also related to clay mineralogy by a simple linear regression analysis of percent shrinkage on percent montmorillonite in the total sample. Montmorillonite was chosen since it is subject to volume change whereas the other clay minerals are not (Grim, 1968). This yielded an r^2 value of only 0.22 and the regression was not significant at the 0.05 level. According to Baver, _et al._, (1972) this relationship should be more obvious than found here. The material being dealt with here in this instance, however, is generally not soil and this may explain the poor correlation since it may behave differently.

5.16 DRYING RATES

Daily moisture samples were taken after one rainfall event until the beginning of the next to determine the rate of desiccation of the material. The data as a function of time (Fig. 5.7) generally show an

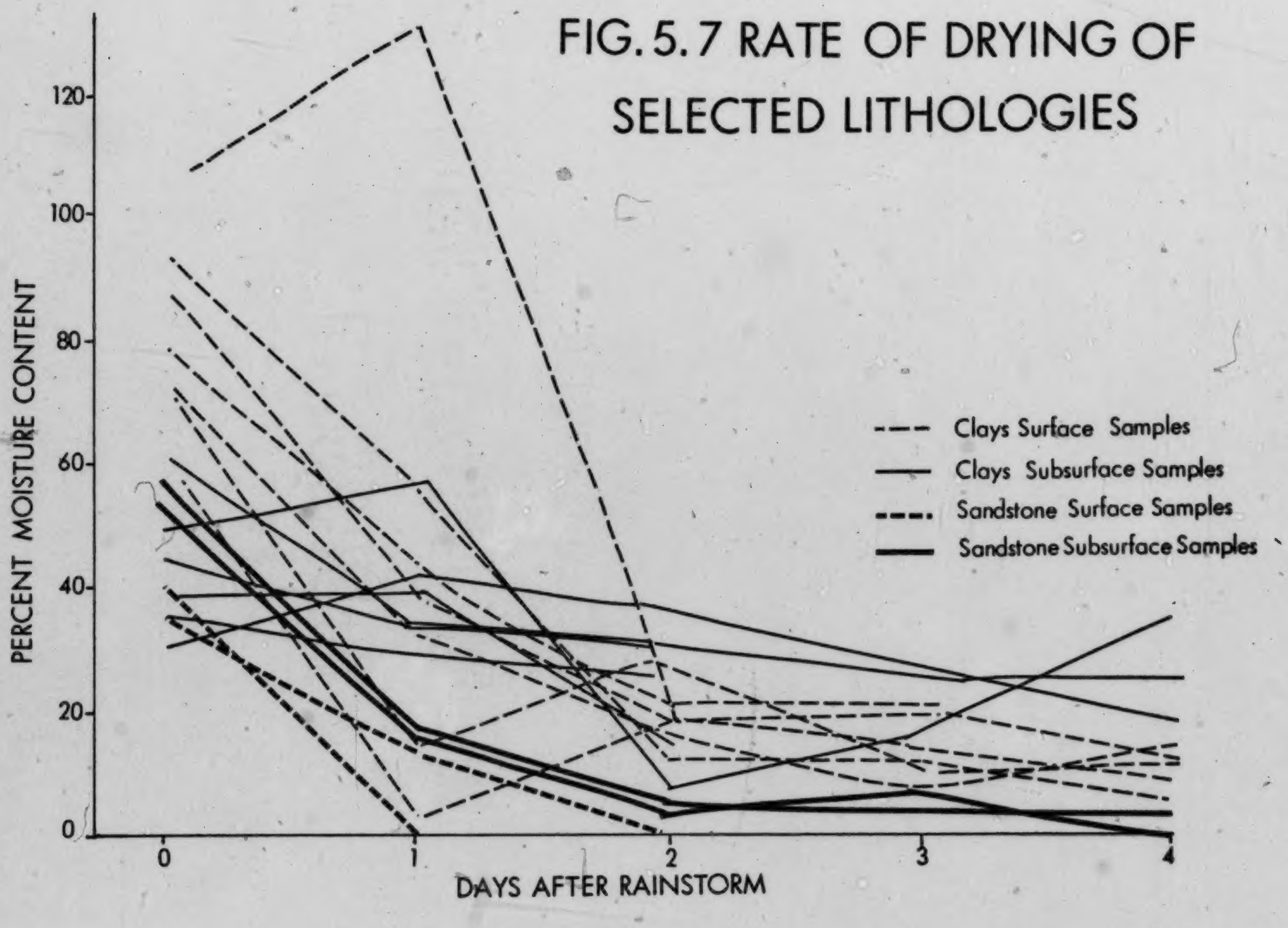


FIG. 5.7 RATE OF DRYING OF SELECTED LITHOLOGIES

initially rapid rate of moisture loss which after four days (the beginning of the next storm) has decreased 60 to 100 percent. Thus, the assumption that the material is "dry" prior to each storm is probably valid. Increases in moisture content shown on these curves are probably due to intra-site variation of moisture content. Subsurface samples appear to be drying at a faster rate than surface samples, contrary to that which is normally experienced in a soil. Corte and Higashi (1964) point out that desiccation cracks are initiated just below the surface which may be the result of this subsurface drying. Also, evaporation may occur in a swelling soil against a moisture gradient due to overburden pressure (Philip, 1971; Sposito, 1973).

5.17 CRACKING EXPERIMENTS

Previous analysis of data from unpublished cracking experiments carried out by Campbell (1968) is presented here and will be integrated with the present study. Data from these experiments appears in Appendix II.

In order to investigate relationships between length of cracking and texture of materials, regression analyses were performed with length of cracking as the dependent variable and percent sand, percent silt and percent clay as alternate independent variables. It was found that percent clay best explains the length of cracking at the 0.05 level (Table 1, Appendix II). It may be concluded from these equations that clay is the main determinant in crack propogation which later in the process may partially be determined by percent silt and still later by percent sand. In all cases, however, r^2 values are low indicating a low level of explanation of length of cracking from these variables. Crack densities range from 0.22 cm/cm^2 to 1.50 cm/cm^2 with a mean of 1.03 cm/cm^2 and a standard deviation of 0.42.

Curves of crack development and water loss through time show an initially rapid development of cracking and loss of water which decreases after about 24 hours (Fig. 1, Appendix II). This laboratory situation is very similar to that found in the field (Section 5.17).

Footnote:

[1] Forsen (pers. comm.) stated that one storm in particular during the summer was extremely unusual in that it continued over a three day period with very low intensity rain rather than the more common sudden high intensity storm of short duration.

CHAPTER VI

DISCUSSION AND CONCLUSIONS

6.1 IMPLICATIONS OF THE RESULTS

Although levels of explanation in the analyses relating moisture contributed from rain to rainfall amount are low, a number of the equations were significant indicating that, although it was not possible to quantify precisely the linear relation, at least it is shown that some relation does exist. In interpretation of field results, due to the variability and complexity of natural processes, the researcher perhaps looks for too high a degree of explanation from simple variables. It should be realised that the method of analysis is not an end in itself, but rather points towards possible explanations (Smith, 1958).

Lithology of the materials appeared to offer little explanation in terms of explaining various moisture contents between different lithologies. This indicates that traditional concepts regarding texture class of a soil and moisture levels are inapplicable in these materials. The materials examined here are not of course "soil" but rather weathered bedrock (except for the two vegetated soil plots). Since the texture of this material seems unrelated to moisture holding capacity then it must be the surface structure of the material which is influencing moisture characteristics and therefore runoff production. Being a fixed variable throughout the profile, texture does not explain differing permeabilities which may be present. The structure is more significant in this regard (Arnett, 1976).

For practical purposes using the normally accepted textural classification for soils on these materials is inappropriate. Material which plots as a sandy clay loam and a silt loam is actually rock with a weathered layer 2-3 mm thick. An unsuspecting agriculturalist would be most surprised on encountering this material in the field.

Taking the best regression equations (multiple regressions of moisture content versus rainfall amount, intensity, and duration grouped according to lithology) comparisons can be made between amounts of shrinkage and clay mineralogy for the same lithologies. The highest r^2 value (0.92) is from a lithology showing moderately high amounts of montmorillonite (34 percent of the total sample) with a high degree of shrinkage (46 percent). Where the highest amounts of montmorillonite are present, however, the degree of explanation for the regression is low (r^2 is 0.30). That is to say, that high amounts of montmorillonite (and associated high degree of shrinkage) appear to be associated with extremely variable moisture contents for given rainfall characteristics. Conversely, low amounts of montmorillonite present in the lithology appear to be associated with relatively precise indications of moisture content from rainfall characteristics. This observation suggests that moisture contents for lithologies containing large amounts of montmorillonite may be directly dependent on the sorption rate of montmorillonite (since rainfall characteristics appear to have little to do with it then it must be some characteristic of the material not measured, such as the rate of sorption). Since this is a known function (Grim, 1968; Mesri and Choi, 1977) with set values over time, then the total moisture content must be some function (though not direct) of the rate of application of rain, i.e., the intensity of rain. A useful analogy to this is a sponge capable of absorbing a large quantity of water. If water is applied slowly then the sponge can soak up its capacity. If, however, the water is applied very rapidly, then much of the water will spill off the sponge because it did not have time to soak in. A comparable analogy for the Hortonian runoff model is a sand or gravel column. If the water is applied more rapidly than it can flow through the column then runoff will occur. The difference between Horton's (1945) concept and the situation suggested here, is that whereas Horton (1945) suggests runoff is a direct function of rainfall intensity, it is suggested here that runoff is a function of rainfall intensity and rate of sorption of the dominant clay mineral, the rate of sorption rather than the rate of infiltration being the limiting factor. Smiles (1974) shows that infiltration into a saturated clay is essentially a sorption phenomenon which would tend to support this argument. The soils, however, do not appear to conform to this generalisation. They show relatively high, or dominant illite content, corresponding low shrinkage values and also low r^2 values. The soil may be behaving more akin to the partial area model in which case rainfall may not have been great enough to initiate runoff. Specific surface for the soils is quite low, but they are fairly sandy, which would imply that water should be able to flow as throughflow in the soil relatively easily.

Some of the materials' hydrologic properties could be tentatively explained through investigations of relationships between moisture contents and texture, specific surface and degree of shrinkage. It has been considered that both composition and structure of the material are responsible for runoff generation in semiarid areas (Peel, 1975; Bryan, et al., 1978). Since little is explained in the present study by the composition of the material then the importance of its structure should be considered. This is extremely complex, and quantitative relationships are difficult to obtain, if not impossible, for the simple approach taken here and consequently field observation of natural processes have to be relied on to a large extent.

Using percent shrinkage data, the area of cracks for a given lithology was calculated. For 100 cm^2 the area of cracks would range from 7.5 cm^2 for soil to 50.5 cm^2 for grey clay. This means that potentially (neglecting adhesion of particles to each other) up to 50 percent of the surface area of grey clay is subject to direct "micro-channel" precipitation at the beginning of a storm. The micro-channel area will decrease over time according to the rate of swelling of the material (Sposito, 1973; Mesri and Choi, 1977). This means that the total amount of moisture able to be taken up by the desiccated "popcorn" terrain type is dependent on rate of swelling as well as rate of sorption and rainfall intensity and amount. If it is assumed that rate of shrinkage is approximately equal to the rate of swelling of a clay (Sposito, 1973) then approximate swelling rates may be calculated from the cracking experiments previously conducted (Appendix II). For grey clay it can be shown that 25 hours must pass before maximum swell has been achieved and the surface is sealed. This is in agreement with times found by Mesri and Choi (1977) for the similar Bearpaw formation. Also, the only time when the surface appeared to be sealed in the field, was after a long (36 hour) gentle rain.

Rainfall intensity does become important with regard to aggregate stability. If aggregates are easily destroyed they may quickly be washed into desiccation cracks and plug the crack reducing further direct water entry. This may significantly speed up the "rate of sealing" of the surface. From the stability tests carried out the most resistant materials seem to be rilled sandstone, white clay and light grey clay. The least resistant are reddish clays and yellow bentonite with the least resistant of all being silty reddish clay. Moderately resistant materials are grey clay and rilled sandstone with lower clay contents. These data show wide variation in separate measurements of the same material and in general the method does not appear to be reproducible unless large sample sizes are used. Continuing with the example of grey clay then, with moderate resistance to raindrop impact, it is reasonable to assume cracks will remain open and clear of sloughed material for a relatively moderate length of time, compared to those materials showing high resistance (cracks remain clear longer) and in those materials with low resistance cracks would be filled relatively soon and puddling of the surface would occur earlier. The rain intensity would determine the absolute times involved since high intensities provide higher energy and greater drop impact destroying aggregates more easily and vice versa. Relative differences between the materials should, however, remain the same.

The angle of incidence of the rainfall also affects the amount of water directly entering the cracks. Vertical rainfall would allow maximum water entry, whereas rainfall driving across the crack at low angles would allow only a relatively small amount of direct water entry since the wall of the crack on the windward side would shelter much of the crack from direct water entry. Angle of ground slope would also be important in this context. Direct water entry into the cracks may therefore occur while the cell (section of material bounded by desiccation cracks) interior is still in a virtually dry state. Only the crack edges and upper surface may be wetted.

Once water has entered the crack, then it must have an outlet or else water will build up in the crack and then flood over the surface. Consequently, the hydraulic conductivity of the subsurface becomes important. In the desiccated "popcorn" material there exists a dense crust which is impermeable except for structural cracks leading into the shards beneath. As the surface popcorn shrinks, it contracts upwards from its base as well as inwards from each side leaving often quite large gaps (3 cm were measured in some instances) between the popcorn surface and the crust (Fig. 6.1). The"sponginess" of the surface when walking over it in a dry state may be attributed to this phenomenon. The result is a subsurface level of disconnected but extensive flow paths with "unlimited" conductivity provided there is an outlet. Outlets are provided by the structural cracks in the crust through which water may rapidly percolate. Water may also simply percolate through the popcorn base to emerge as throughflow at the base of slopes. When the shards beneath the crust come into contact with water they immediately slake (Bryan, et al., 1978) and thus, if not already in existence, a subsurface pipe may begin.

Fig. 6.1. Gap produced by shrinkage of the "popcorn" layer upwards. The pen is standing in the gap (the surface is removed) to a depth of 2.5 cm.

If water is unable to pass rapidly enough through the crustal cracks, because they are too narrow, once a head of water is established there may be enough energy for the water to force its way into the shard layer thus allowing rapid drainage of the water which has built up (Barendregt and Ongley, 1977). Where water does not penetrate to the shard layer, then throughflow may occur immediately above the crust surface (Fig. 6.2). In either case water flows out at the base of slopes on the pediments as micro-pipe flow or throughflow. An interesting observation was made on Plot One when a footprint in wet clay left a small depression in an otherwise uniform slope. After the next storm event a tiny alluvial fan had built up from micro-pipes issuing water and sediment at the break in slope (Fig. 6.3).

These concepts are shown schematically in Fig. 6.4, presented as a runoff generation model for the "popcorn" terrain type. During the first stage, rain begins and enters the profile by direct micro-channel precipitation and flow begins along the crust surface and through structural cracks penetrates to the shard layer to join pipe networks. The edges of the cells only, are wetted at this stage. As rainfall proceeds, the cells absorb more moisture and begin to swell. The surface begins to swell first since it is receiving the most water. Because of the swelling, the crack begins to close, but closure occurs from the top first, since it is the wettest. Flow continues through the system. After further swelling, stage three is reached where the crack is completely closed. Water now enters only by infiltration through the material and sorption processes of the clay. As water content of the cells increase, pore water pressure increases causing the expulsion of water through the

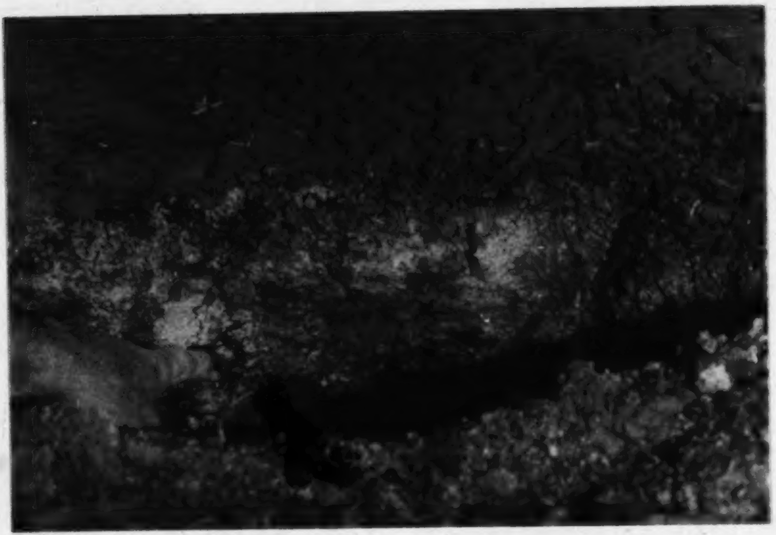
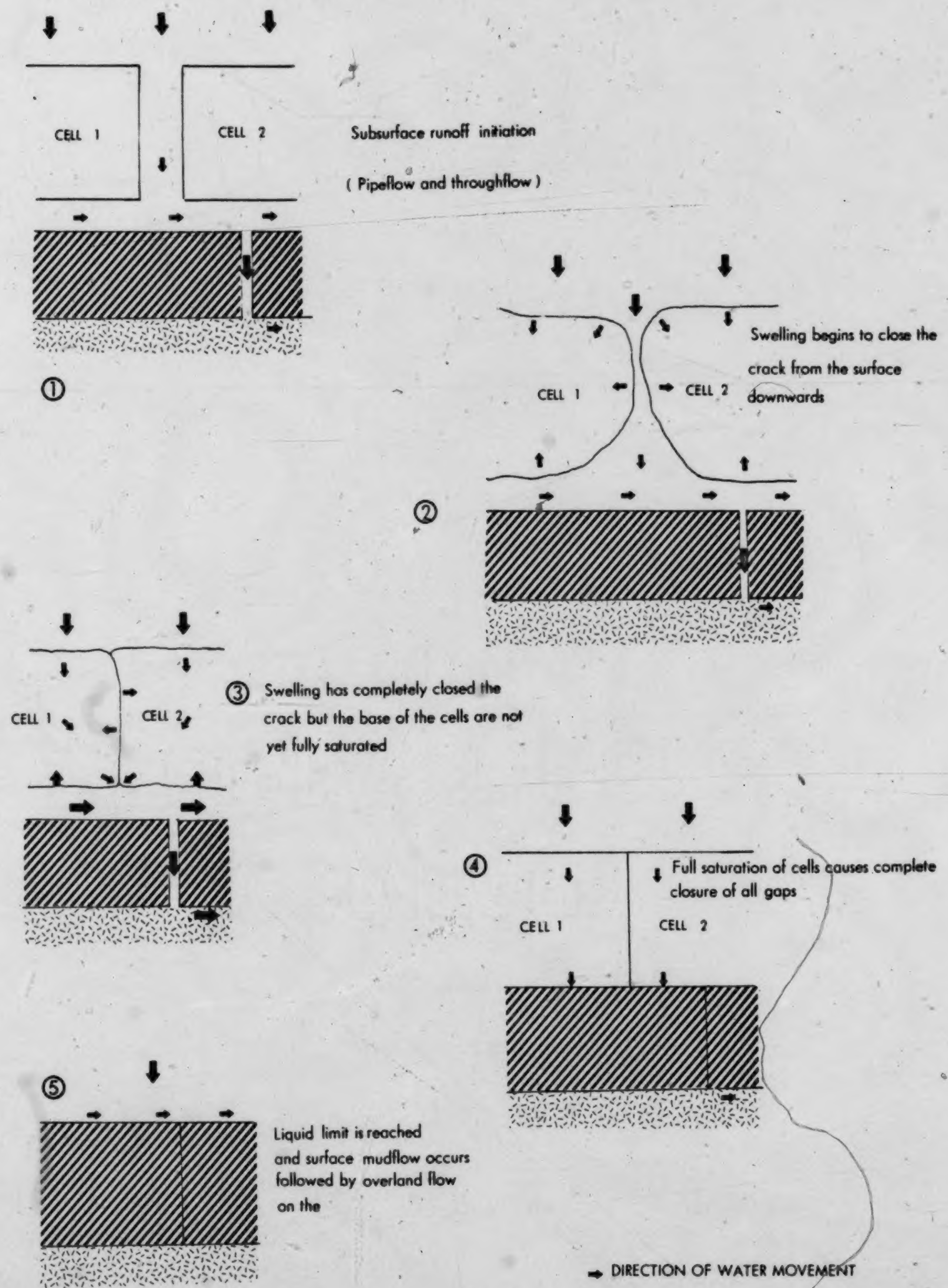


Fig. 6.2. Throughflow (the dark area in photograph) above the crust layer in shale.

Fig. 6.3. Small alluvial fan initiated by a footprint with micropiping evident from the flow lines of lighter coloured sediment.

FIG. 6.4 SCHEMATIC REPRESENTATION OF RUNOFF PROCESS ON POPCORN TERRAIN TYPE


CELL 1
CELL 2
①
CELL 1
CELL 2
②
CELL 1
CELL 2
③
④
CELL 1
CELL 2
⑤

① Subsurface runoff initiation
(Pipeflow and throughflow)

② Swelling begins to close the crack from the surface downwards

③ Swelling has completely closed the crack but the base of the cells are not yet fully saturated

④ Full saturation of cells causes complete closure of all gaps

⑤ Liquid limit is reached and surface mudflow occurs followed by overland flow on the

→ DIRECTION OF WATER MOVEMENT

wettest parts of the base and continued flow along the crust surface is maintained. The base of the cells are not yet wet enough to cause maximum swelling. Eventually, however, with continued infiltration and sorption the gap between the crust and "popcorn" layer is closed due to the cell base swelling. The degree of swell is directly proportional to the volume of water taken up by the clay (Smiles, 1974). The presence of gypsum crystals in some areas may decrease the amount of swell in some localised areas (Bridge and Turry, 1973). Some flow will still occur through the cracks in the crust due to increased pore water pressure. However, once the liquid limit (Table 5.7) of the "popcorn" material is reached, surface flow may occur as a small mudflow sliding over the crust which acts as a basal slippage plane (Grim, 1968; Bryan, _et_. _al_., 1978). Direct rainfall on the crust then causes overland flow since the crust is impermeable (except for the structural cracks). Pipeflow may still continue due to the presence of structural cracks. Thus the final stage of the runoff process is a small mudflow enhanced by overland flow immediately following the mudflow, occurring simultaneously with the subsurface pipeflow. The final stage is probably reached after about 6 mm of rain, but this would depend on the slope and this figure is for a steep slope (Bryan, _et_ _al_., 1978).

This sequence of events was observed when attempting to measure rates of vertical infiltration using infiltration rings which proved impossible since lateral flow could not be contained as the crust was too dense to penetrate with the rings. Read (cited in Collis-George, 1977) used a small pile driver to force rings down to depths of 30 cm (Read, pers. comm.) but this would surely disturb the soil to a great extent, which would make measurements non-representative of the true situation.

In summary, the runoff (R.O.) from the "popcorn" terrain type may be stated as

R.O. = f(width of crack, rate of swelling, aggregate stability, intensity of rain, duration of rain, angle of incidence, of rain, rate of infiltration into the cells, rate of sorption of the dominant clay mineral, hydraulic conductivity of the subsurface, angle of slope ...)

This function is obviously extremely complex and requires much detailed investigation of the behaviour of this material in situ. Disturbance of the material would be too great to enable useful laboratory experiments in the investigation of this function. Furthermore, it is obvious that this terrain type does not behave according to the Horton (1945) runoff model. The relationship between rainfall intensity and infiltration rate in this case is not simplistic and surface overland flow does not occur until the final stage of the process. On low angle "popcorn" terrain types, stage five may be less frequently reached. Flow occurs mainly as subsurface pipeflow and throughflow. Where rills are present, flow does occur in the rills due to decreased infiltration in the rills from deposition of material on their floor (Bryan, et al., 1978); but here the materials are completely different from the clay cemented sandstone, the surface of which is smooth in comparison to the "popcorn" terrain type. Flow occurs in the rills on the sandstone simply because of the existence of the channels whereas rill flow on the "popcorn" terrain occurs as a consequence of both the presence of the channel network and the decreased infiltration in the channel. Since drainage density tends to reflect terrain permeability (Garner, 1974) and in view of the evidence discussed previously of the number of rills on the two surfaces, it is obvious that the "popcorn" terrain type has greater permeability than the clay cemented sandstone.

Surface runoff on the clay cemented rilled sandstone and micro-pediment terrain types is practically instantaneous (Bryan, et al., 1978). It seems reasonable to suggest that Horton's (1945) infiltration-excess overland flow model would apply to these terrain types. Little relation of moisture content to characteristics of the material could be found for the rilled sandstone. Depth of wetting on sandstone, however, was only in the order of 2-3 mm (depth of weathered layer) after each storm so that considerable dehydration may have taken place before moisture samples could be taken. This may also have affected moisture contents on other terrain types. Where pipes do occur in sandstone, however, obviously Horton's (1945) model is modified to a combination infiltration-excess or saturation overland flow (Kirkby and Chorley, 1967) and subsurface pipeflow. Since the weathered material is quite sandy, then infiltration rate should not be a limiting factor. In view of this, overland flow that does occur is probably saturation-overland flow (Kirkby and Chorley, 1967). Surface flow occurs so rapidly because the profile to be saturated is only 2-3 mm thick. Beneath this is unweathered, compact bedrock.

The micro-pediments were not examined in detail in this study, but Bryan, et al. (1978) suggest a true Hortonian model for runoff generation. Wetting of the profile takes place but to a limited depth (2-3 mm) and infiltration is limited by compaction which produces an extremely smooth surface on the micro-pediment. Runoff from natural rainfall may occur as a continuous sheet of water moving over the surface. This was observed on one occasion under natural conditions in the field, (Fig. 2. 10).

The soils runoff process seems to be less obvious and their behaviour can only be surmised. As discussed previously the storm events investigated here may be too small to initiate runoff from these surfaces. At no time was surface flow observed on these terrains except where it was on a bed of material deposited over the vegetation. These "beds" were presumably produced as "flood" deposits from surrounding micro-pediments. Within the soil terrain type, however, is an extremely large soil pipe. On an exposed wall of this pipe, seepage was observed at several horizons in the profile. This indicates that subsurface throughflow almost certainly contributes to flow in this pipe and is at least possible in the soils terrain type. Water should be able to move relatively easily through these soils since they have a relatively sandy texture (loam). Consequently it seems reasonable to suggest a saturation overland flow/throughflow model (Dunne and Black, 1970a; Weyman, 1975) for the soil terrain type.

6.2 CONCLUSIONS

The conclusions from this study are set out below:

1. Lithology and mineralogy are highly variable within the basin which presents a number of problems; apart from difficulty in mapping, extreme care must be taken when transposing results to other areas

even within the same locality. Lithology and mineralogy changes as rapidly horizontally as it does vertically with extensive interfingering of deposits occurring. This has implications relating to the scale of the study. Runoff from a large scale (basin) study may not be influenced by lithology and mineralogy to the extent that small scale (plot) processes are, since an averaging out of variations probably occurs due to the longer times involved for processes at the larger scale. This has not been fully recognised in the past and is of significance to future studies, particularly sprinkling experiments, since these are usually on a small scale and may not relate, in absolute terms, to the larger scale, basin runoff.

2. Moisture content of the "soil" is linearly related to rainfall amount below a threshold amount of rain (variable according to lithology but probably not greater than 12 mm). This implies that before any appreciable runoff occurs in the basin, up to 12 mm of rain may fall and be infiltrated into the "soil". Consequently, in view of the possibilities for throughflow and subsurface movement of the infiltrated water, it seems reasonable to apply the partial area concept to runoff generation in the semiarid environment as well as the humid environment.
3. Aspect has some, at present unquantified, affect on the moisture contents of the materials. Micro-climatic variables appear to be significant to the hydrology of these areas. The findings of this study compare with suggestions made by Beaty (1975b). A seasonal cycle of relative permeable-impermeable surfaces such as that suggested by

Schumm and Lusby (1963) and Schumm (1964b) may be evident here, except that it is restricted to the lee side of slopes where snow accumulation occurs (Beaty, 1975b). This would be modified, however, since rainbeat occurs mainly on the windward side of the slopes. Thus a cycle of permeable-impermeable surfaces may not occur at all. Rather, in winter the lee slopes may be active due to freeze thaw activity and shrink-swell phenomena, while in summer the windward slopes may be active due to shrink-swell phenomena and rainbeat. It is probable the major role of snow melt is to flush out accumulated debris in the spring causing high sediment yields during the spring melt season. Contrary to that which is normally understood for aspect relationships insolation in the case of the Steveville badlands probably has only a secondary effect related to drying rates of the material, but here also wind is important.

4. A relationship between textural class and other lithologic variables to moisture content could not be defined. This suggests that traditional soils techniques may need to be modified when dealing with weathered bedrock of this nature. Replication of field conditions were impossible in the laboratory due to the delicate, but hydrologically significant, microstructures of the surficial material.
5. Structure of the material is more important in terms of runoff generation than lithology. Structure may or may not provide pathways for subsurface flow.
6. Micro-channel precipitation is an important factor in regard to moisture contents and runoff generation of desiccated materials resulting

in a partial area contribution to runoff on a micro-scale at surface and subsurface levels.

7. Surface runoff is more dominant on the sandstone slopes whereas subsurface runoff is more dominant on the shale slopes. Where sandstone rests on a shale unit, the runoff may begin on the surface, flow along subsurface paths in the shale, and reappear at the surface again when the next sandstone unit is encountered. Thus, runoff may follow a rather complex route to the basin mouth and may in fact never get there.
8. A number of different runoff generation models are proposed depending on the terrain type, emphasising the variety and complexity of the processes active in badland areas. Horton's (1945) runoff model does not apply generally to the clay badlands as has been previously suggested. Rather, it is restricted to the micro-pediment surfaces. A model incorporating micro-channel precipitation and rates of swelling is suggested for the "popcorn" shale surfaces while a saturation-overland flow model is suggested for a sandstone surface. Previously, such partial area models were applied to basin studies, whereas here they apply to micro-scale features such that several models are represented in one basin.
9. In order to verify the runoff models proposed here, a field approach is required, since laboratory conditions cannot replicate the complex field micro-structures which are significant in the runoff process.
10. Care should be taken in treating badlands as "geomorphic field laboratories." Processes concerned with runoff are different here than in

humid areas and involve unique processes different from other more climatically and lithologically uniform areas. Furthermore, extrapolation of results from one area of badlands to another should be done with extreme caution, since minor variations in lithology and climate (see 1.) may alter enormously the rate of or even the characteristics of the process in question. The sensitivity of badlands environments to these types of variables has not been fully recognised and much research is required particularly investigating the effects of microclimate on the hydrologic balance of the small and large scale areas. Some investigation of the nature of the effect that small scale processes discussed here have at the drainage basin scale is also required.

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APPENDIX I

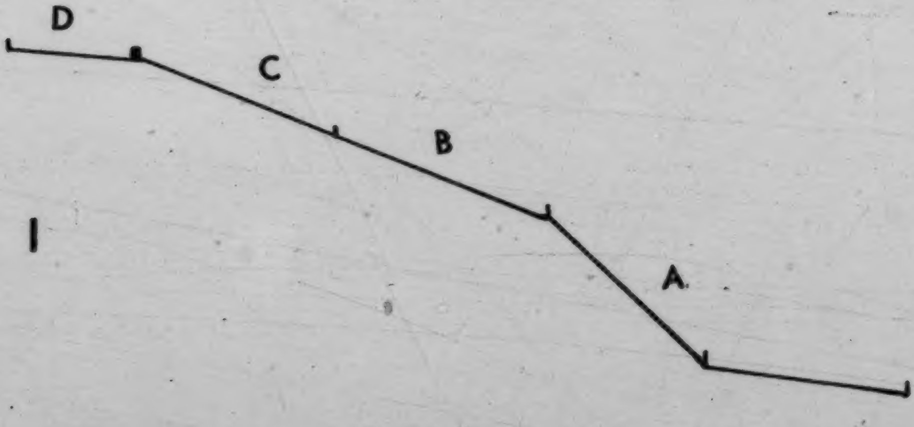
DETAILS OF INDIVIDUAL SAMPLING SITES

1. Photographs of each plot with vertical (downslope) profiles.

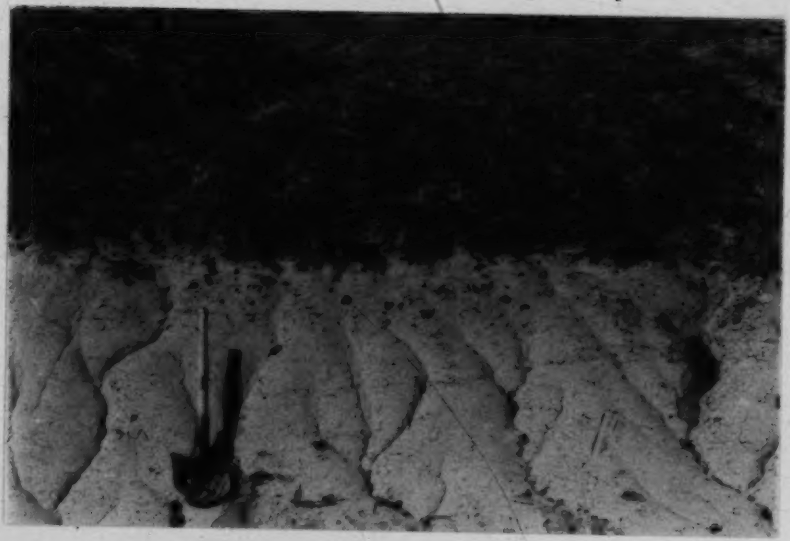
    Horizontal scale 1 cm = 33 cm

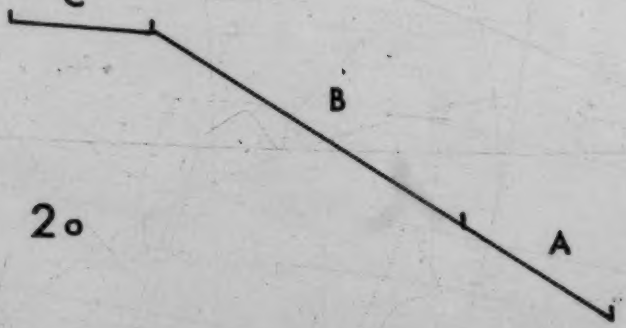
    Vertical scale 1 cm = 33 cm

    Vertical exaggeration = 1



1





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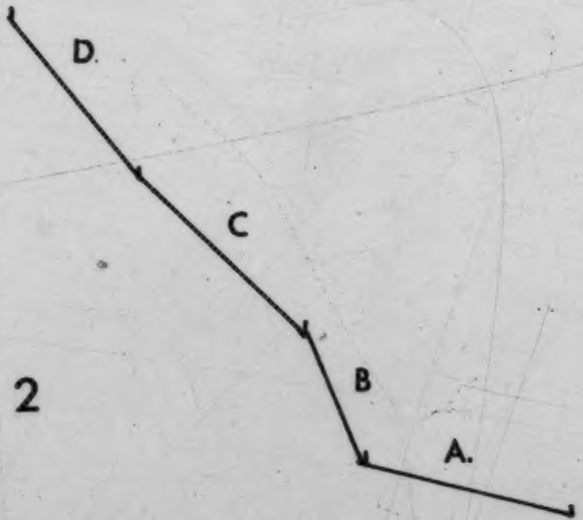

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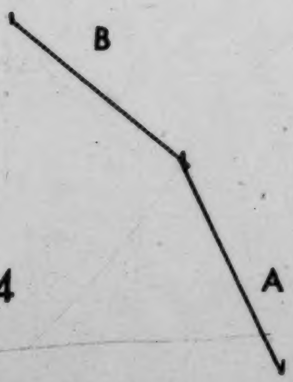

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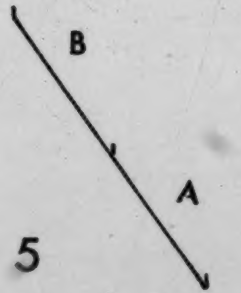

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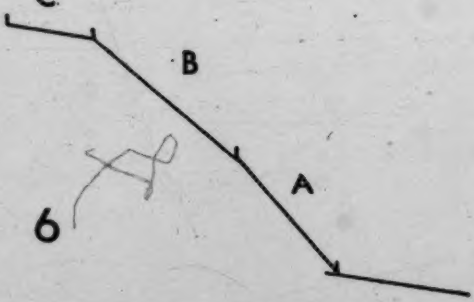

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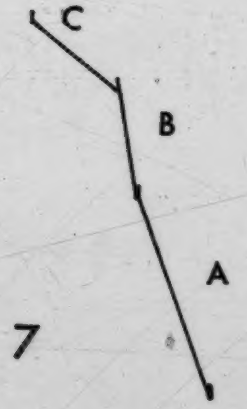

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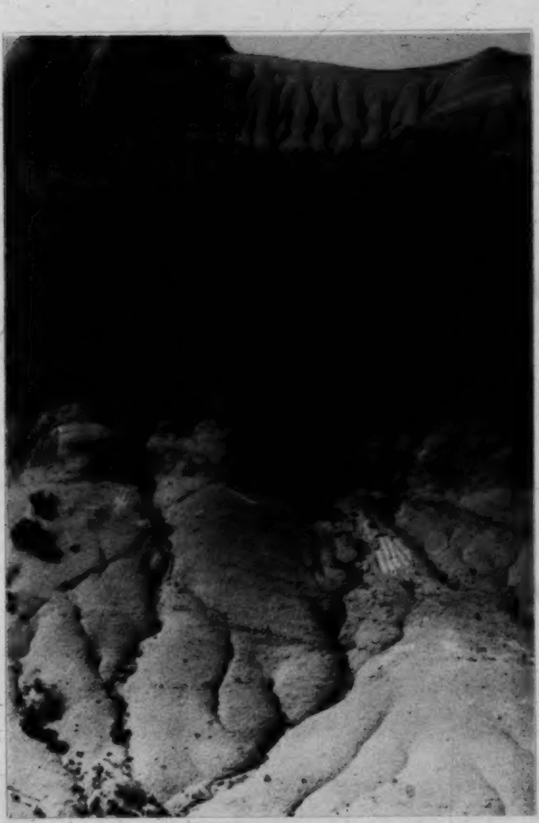

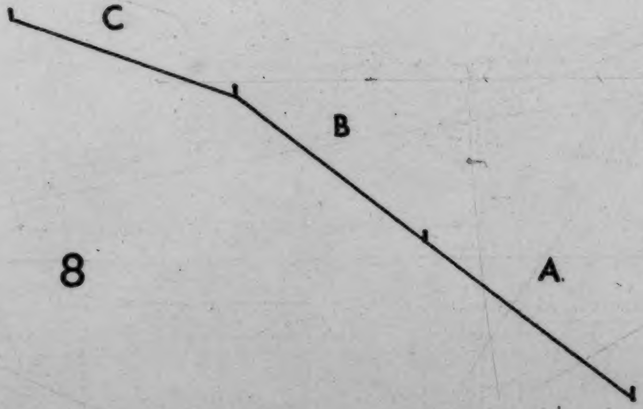

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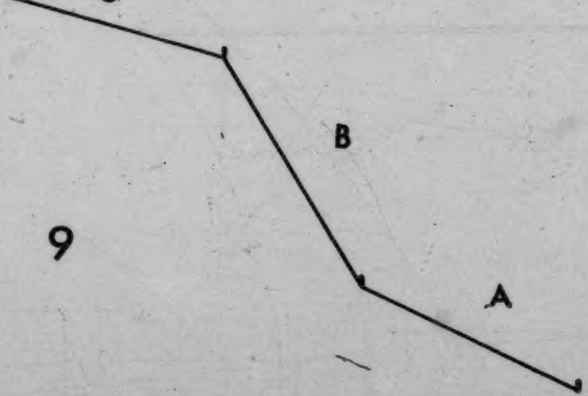

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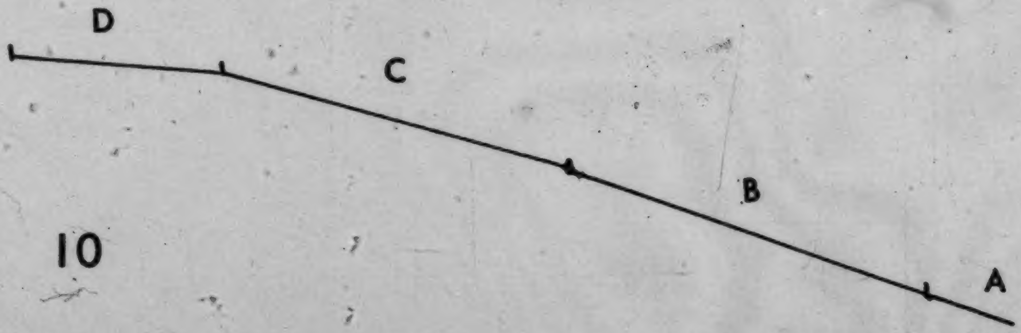

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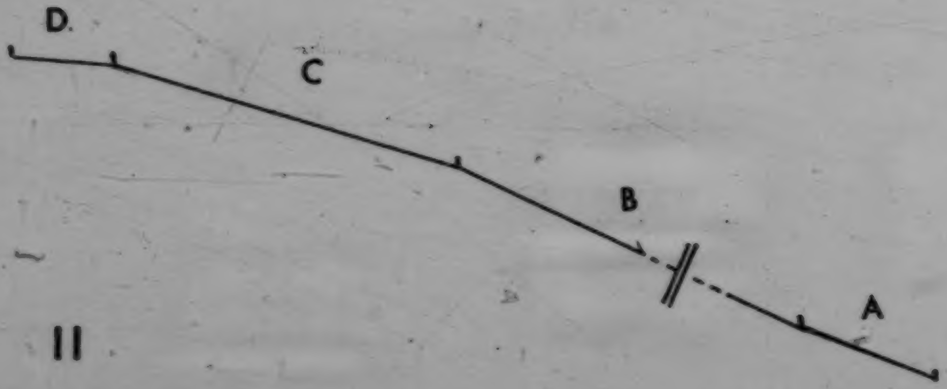


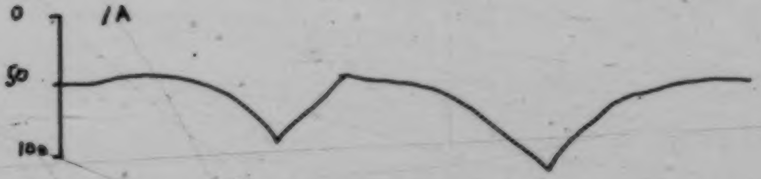
11

2. Horizontal (across slope) plot profiles.

Horizontal scale 1 cm = 10 cm

Vertical scale 1 cm = 5 cm

Vertical exaggeration = 2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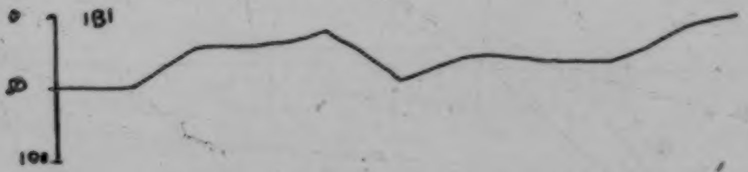
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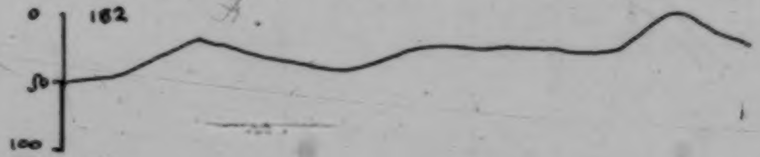
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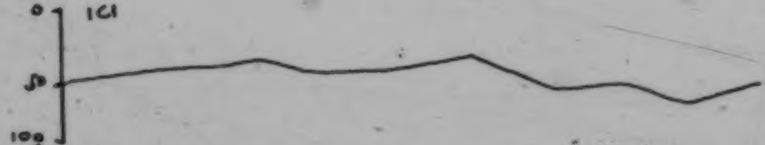
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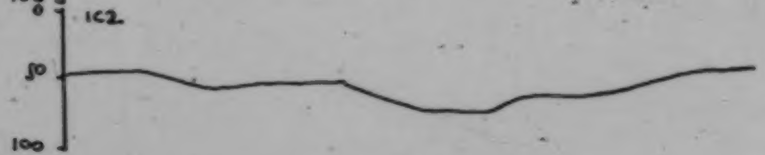
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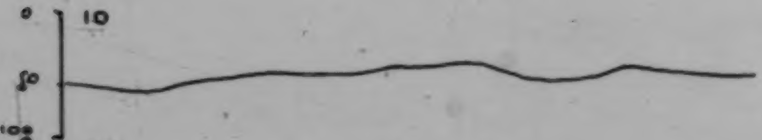
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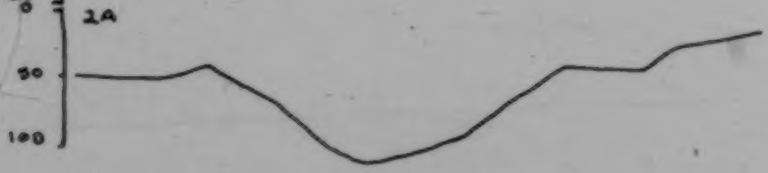
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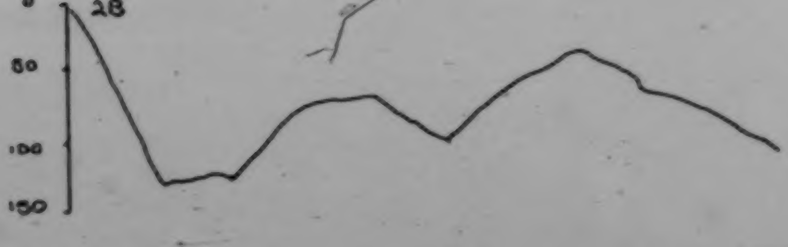
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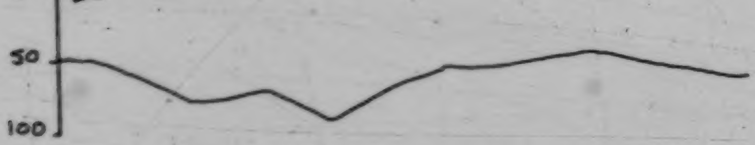
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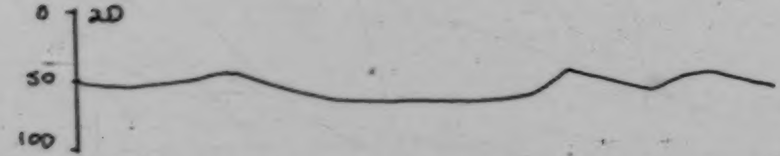
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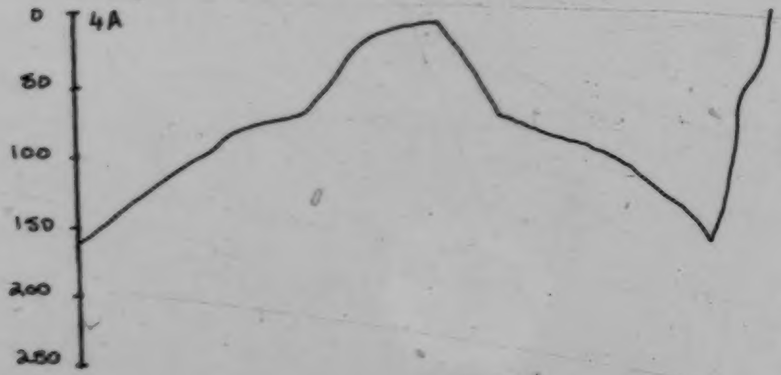
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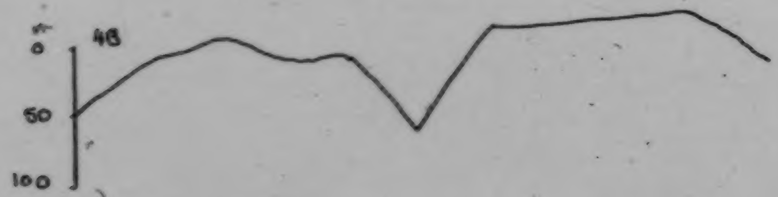
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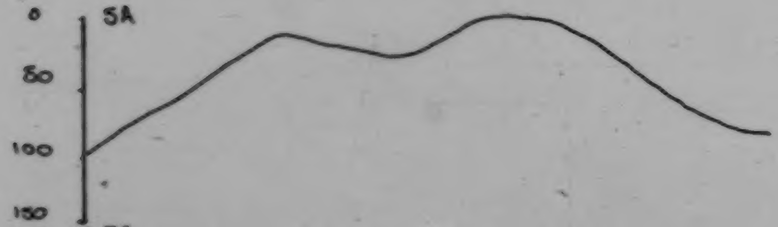
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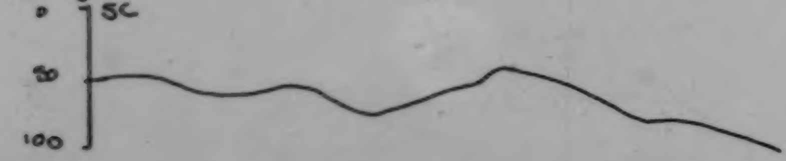
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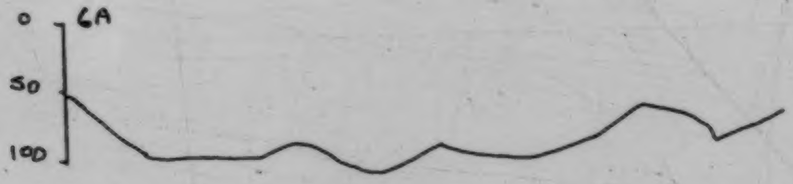
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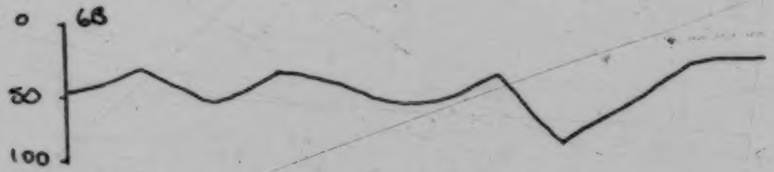
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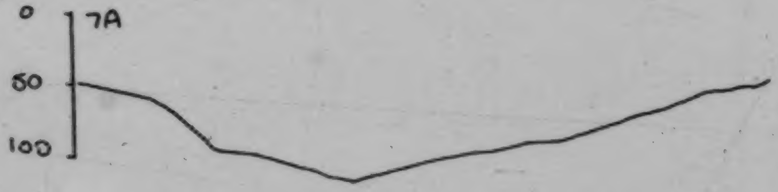
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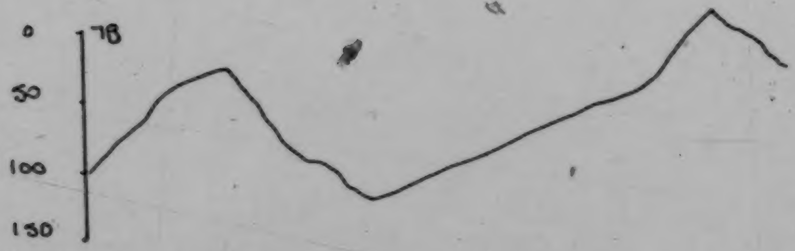
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100
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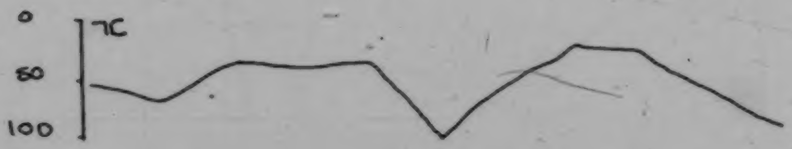
7C
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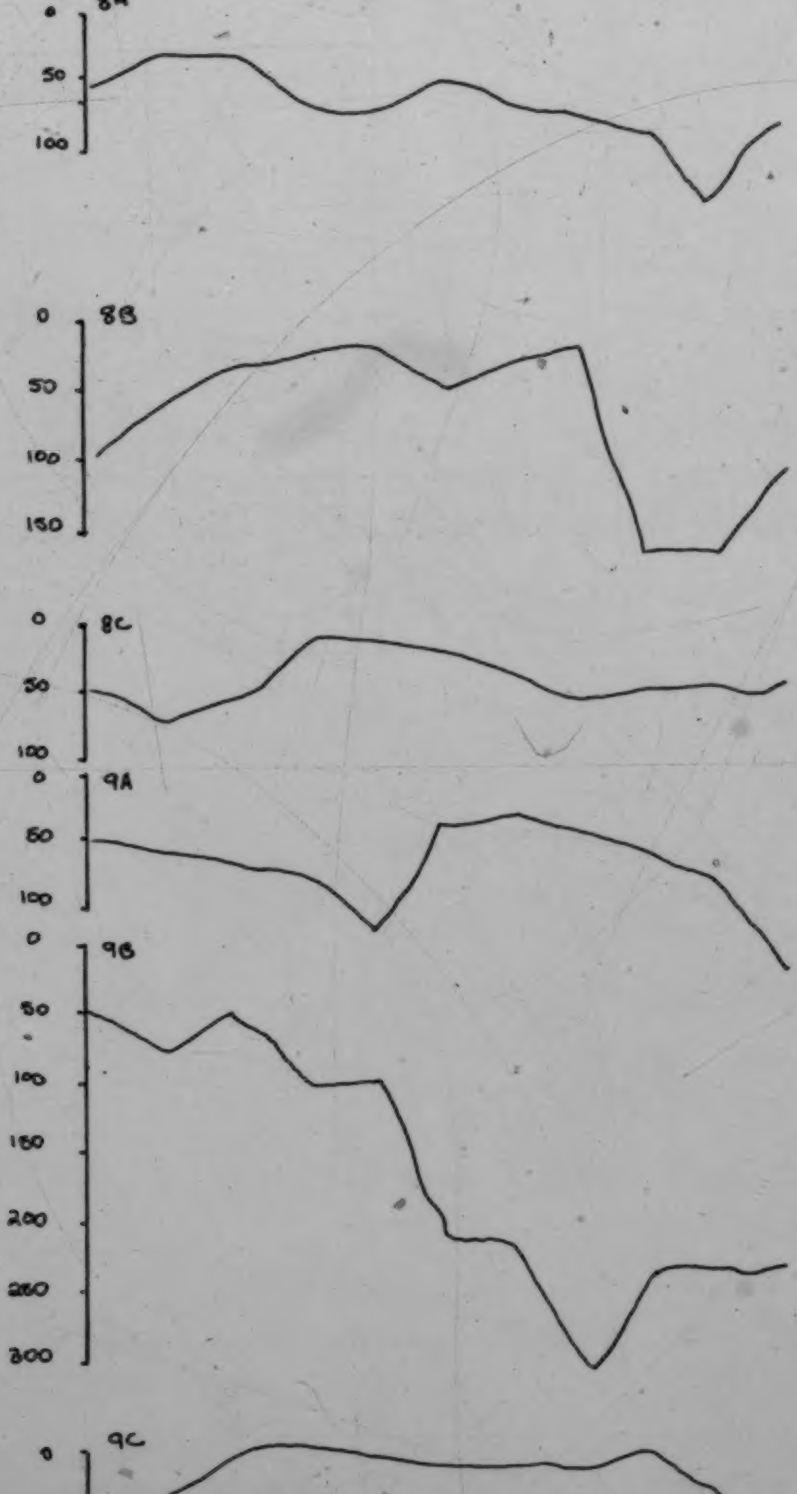
8A
0
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100
8B
0
50
100
150
8C
0
50
100
9A
0
50
100
9B
0
50
100
150
200
250
300
9C
0
50
100

Table 1. Plot and sample sites with Munsell soil colours, average crack width and average cell size for each site.

Plot	Sample Site	Colour	Crack Width (mm)	Cell Size (mm^2)
1	A	5Y 5/2 greyish yellow	2.4	219
	B1	5Y 7/2 light grey	1.6	218
	B2	10YR 5/3 dull yellowish brown	2.2	6308
	C1	7.5YR 6/6 orange	2.2	1302
	C2	7.5YR 6/3 dull brown	2.3	306
	D	10YR 6/2 greyish yellow brown	1.5	450
2	A	2.5YR 5/1 yellowish grey	0.65	615
	B	5Y 5/1 grey	1.1	535
	C	7.5Y 6/1 grey	1.2	210
	D	2.5Y 6/3 dull yellow	3.8	623
4	A	2.5Y 7/1 light grey	2.4	865
	B	2.5Y 6/1 yellowish grey	2.1	3068
5	A	2.5Y 6/1 yellowish grey	3.6	385
	B	2.5Y 6/2 greyish yellow	7.4	191
6	A	10Y 8/1 light grey	1.2	338
	B	7.5Y 8/1 light grey	0.2	227
7	A	5Y 8/3 pale yellow	0.8	225
	B	5Y 8/1 light grey	0.8	291
	C	5Y 7/1 light grey	4.2	1072
8	A	2.5GY 7/1 light olive grey	0	0
	B	7.5Y 7/3 light yellow	6.2	855
	C	5Y 6/4 olive yellow	4.5	873

Table 1. (Cont.).

Plot	Sample Site	Colour	Crack Width (mm)	Cell Size (mm^2)
9	A	7.5Y 7/1 light grey	0	0
	B	5Y 7/4 light yellow	0.8	443
	C	5Y 7/2 light grey	5.7	461
10	A	2.5Y 5/3 yellowish grey*		
	B	2.5Y 5/2 dark greyishy yellow		
	C	2.5Y 5/3 yellowish grey		
	D	2.5Y 6/1 yellow grey		
11	A	5Y 5/2 greyish olive*		
	B			
	C	2.5Y 4/3 olive brown		
	D	2.5Y 5/2 dark greyish yellow		

*the soils do not have desiccation cracks comparable to the other terrains and therefore were not investigated for crack width and cell size.

APPENDIX II

DETAILS OF CRACKING EXPERIMENTS CONDUCTED BY CAMPBELL (1968)

1. Experimental procedure.

   1. Break up soil clumps and crush with mortar and pestle.
   2. Add 100 ml of water with a spray can.
   3. Weigh sample.
   4. Place in oven at 100°F (38°C).

Measurement times were	A	after 30 mins	30*
	B	" 30 mins	60*
	C	" 120 mins	180*
	D	" 180 mins	360*
	E	overnight	1275*
	F	after 375 mins	1650*

* cumulative time increments (minutes)

Table 1. Crack length, particle size and clay mineralogy for samples used in desiccation crack experiments.

Sample	Crack Length	Percent Sand	Percent Silt	Percent Clay	Percent Montmorillonite	Percent Illite	Percent Kaolinite
1	340	15.0	14.5	70.5	80	20	0
2	308	26.3	57.5	16.2	100	0	0
3	245	8.7	31.8	59.5	5	80	15
4	361	7.5	36.1	56.4	15	60	25
5	51	59.5	37.7	2.8	10	70	20
6	275	2.8	52.8	44.4	70	25	5
7	284	18.2	24.4	57.4	5	65	30
8	203	8.9	49.7	41.4	100	0	0
9	252	17.6	28.6	53.8	90	5	5
10	225	15.9	50.0	34.1	70	25	5
11	223	59.2	24.8	16.0	100	0	0
12	162	14.4	46.0	39.6	10	75	15
13	43	31.3	65.1	3.6	70	30	0
14	312	19.8	15.4	64.8	65	30	5

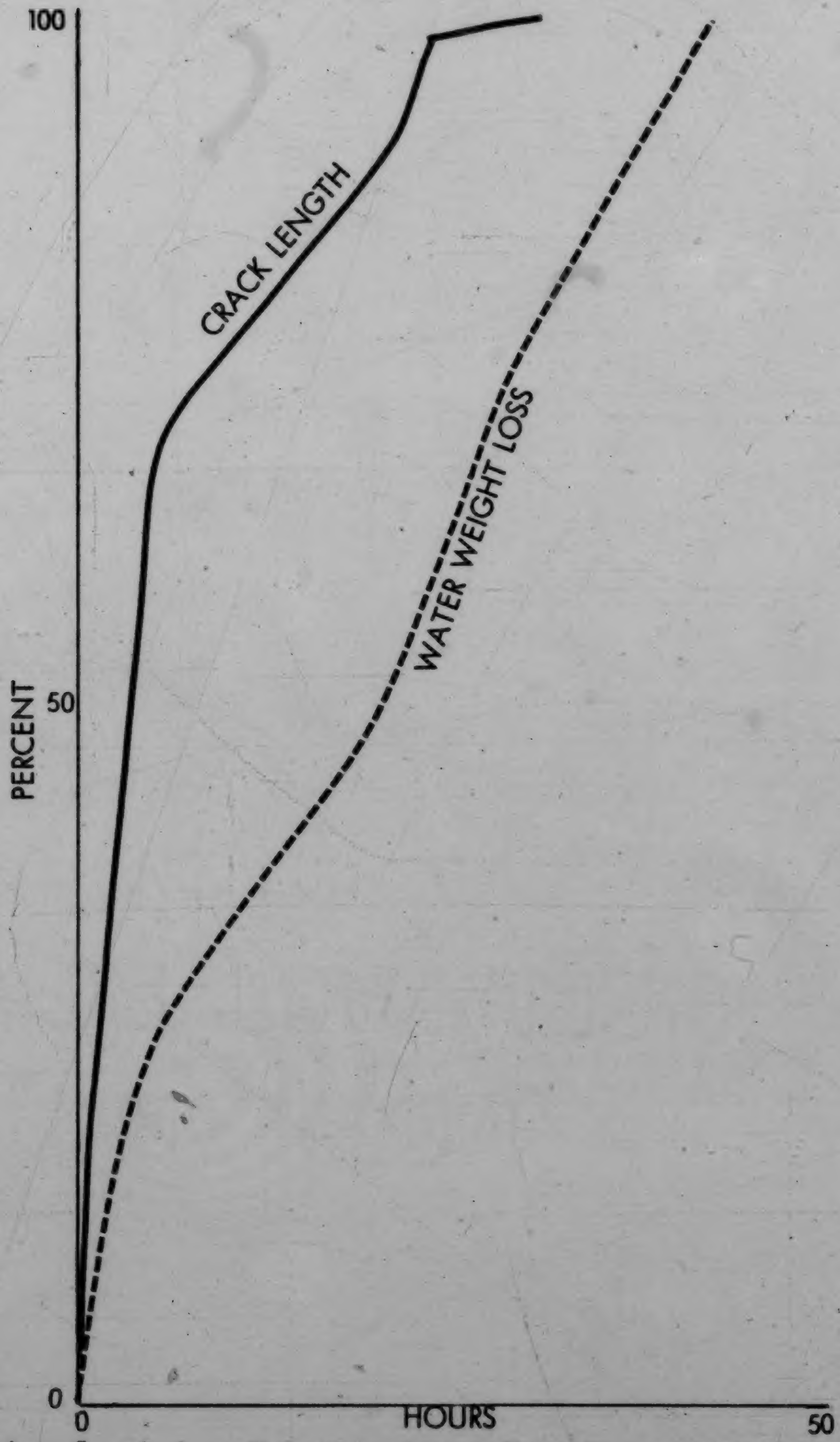

Fig. 1. Crack length and water weight loss over time.